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STRATIGRAPHY OF THE LATE PLEISTOCENE LOESSES OF KANSAS[1]

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State Geological Survey, University of Kansas

ABSTRACT

The post-Yarmouthian Pleistocene of Kansas is characterized by extensive loess sheets which contain buried soils. The loesses and associated fluvial deposits are classed in Kansas as Crete, Loveland, Peoria, and Bignell members of the Sanborn formation. Based on lithologic characteristics, sequence, content of fossil mollusks, and stratigraphic continuity of buried soils, the members are identified throughout their extent in Kansas. Also trustworthy correlations with late Pleistocene deposits in Nebraska and Iowa, including the type Bignell and Loveland loesses, can be made. The Peoria loess of Kansas is shown by faunal evidence to be equivalent to the Peoria (Iowa and Tazewell) loess of central Illinois.

INTRODUCTION

The stratigraphic units most widely observable in outcrop sections in the central United States are silts or loesses of late Pleistocene age. Deposits of massive silts of late Pleistocene age have been described from localities distributed from eastern Colorado to Indiana and from Iowa to Mississippi. In Kansas they form the surficial material of more than one-third the state, especially in the upland area of northern and western Kansas. The establishment of a usable stratigraphic classification and the correlation of Pleistocene units in different parts of the state with one another and with the succession of upper Mississippi Valley glacial and interglacial deposits have been attempted only during the past decade.

It is the purpose of the present paper to describe the stratigraphy of these deposits in Kansas and to discuss the correlation of the several recognized stratigraphic units with described deposits in adjacent areas. Field and laboratory studies of Kansas loess deposits have been supported by the State Geological Survey during much of the last ten years. During the field seasons of 1948 and 1949, several months were devoted by us to state-wide study of the stratigraphy and paleontology of the Kansas loesses. Important contributions to our data were made by members of the co-operative federal and state Ground-Water Division, by Ada Swineford and Norman Plummer of the State Geological Survey, and by James Thorp and others of the Soil Survey Division, Bureau of Plant Industry, Soils, and Agricultural Engineering of the U.S. Department of Agriculture.

Field studies consisted of examination of surface exposures, measurement of stratigraphic sections, collection of lithologic samples, examination of molluscan faunas in place, and the selection of lo-

[1] Manuscript received July 12, 1950.

calities for quantitative sampling of faunas. At many places, including the High Plains of west-central Kansas, exposures are inadequate for detailed study; in such areas samples of loess and faunas and data on buried soils and lithology were obtained by the use of a 3-inch spoon-type soil auger. Methods of lithologic sampling and petrographic study of the Peoria loess are described by Swineford and Frye (1951).

EARLY PLEISTOCENE HISTORY

The Pleistocene geologic history of Kansas is logically divided into two segments of almost equal length, separated by the Yarmouthian interglacial interval. During early Pleistocene time, northeastern Kansas was twice invaded by continental glaciers, and the effects of these glaciers extended throughout the state, reaching far beyond the ice margins. Prior to the first glaciation, Kansas was an area of relatively low relief; in the west an aggradational plain marked the top of fluvial Pliocene sediments, and the eastern third of the state, east of the Flint Hills divide, was an erosion surface crossed by shallow valleys filled with locally derived materials. The two pulsations of continental glaciation caused successive episodes of erosion and sedimentation, accompanied and followed by important modifications of major drainage lines. The record of early Pleistocene time is preserved not only in the glaciated area (about one-twentieth the area of the state) but at localities within the Plains region several hundred miles from the ice border where stream deposits have been preserved.

A prominent erosional unconformity occurs generally between the Ogallala formation, of Pliocene age, and early Pleistocene deposits. The oldest sediments classed as Pleistocene are judged to be related genetically to the advance and retreat of the Nebraskan ice. Alluvial sediments of Nebraskan age occur generally as fillings in valleys, many of which are now abandoned by major drainage, and as basin fillings in a few places. The areas determined to have constituted divides during Nebraskan time do not contain sediments of that age, although it must be admitted that eolian sediments, if once present on these divide areas, would have been particularly vulnerable to erosion.

These early Pleistocene sediments, although generally comparable in lithology to deposits made by continental and Rocky Mountain glaciers, in some places are composed largely of locally derived materials (Fent, 1950*a*, *b*); therefore, it is evident that the cycle of erosion and sedimentation associated with glaciation was operative in drainage basins isolated from direct effects of the ice.

Between Nebraskan and Kansan time, erosion and sedimentation proceeded at a greatly reduced rate, but during Kansan glaciation a cycle of cut and fill, similar to that of Nebraskan time, occurred. The general drainage pattern now existing throughout Kansas is judged to be the product of events of Kansan time, although some stream piracies and valley abandonments initiated then may not have been consummated until later within the Pleistocene. Indeed, some may be still in progress. Deposits of Kansan age, like those of Nebraskan, occur largely in valley areas. Although silt deposits are widely prevalent in the Sappa member of the Meade formation (latest Kansan to earliest Yarmouthian), they generally show evidence of water deposition, and the former divide areas commonly are free of Kansan sediments.

Following the episode of sedimentation that produced the Meade formation,

most of Yarmouthian time was an interval of relatively little erosion or deposition but of extensive soil formation.

CLASSIFICATION OF LATE PLEISTOCENE DEPOSITS

Late Pleistocene deposits in Kansas are classed as belonging to the Sanborn formation, which includes, in the basal part, the Crete sand and gravel member, which is overlain by (1) the Loveland silt member, terminating upward in the prominent Sangamon (Loveland) buried soil; (2) the Peoria silt member, terminating upward in the Brady soil; and (3) the Bignell silt member, terminating upward in modern topsoil. Low terrace surfaces along many valleys of the state are underlain by sediments which may correspond in age to the Peoria and Bignell members. These materials, however, are not included in the Sanborn formation because they are commonly mapped as physiographic units, on the basis of surface form and extent of the terraces, rather than as stratigraphic units. Also, the lithology of the terrace materials is commonly unlike that of the Peoria and Bignell members. Colluvial slope veneers of indeterminate age are also excluded from the Sanborn formation.

Widespread deposits of massive silt in northern and western Kansas were recognized as early as 1866 by Hawn (p. 101), who referred them to the "Bluff formation" along the Republican Valley. Hay (1885, 1893) described upland silts in the High Plains region as the "Plains marl or Tertiary marl." Haworth (1897, p. 276) and Williston (1899, p. 287) considered the extensive silt deposits of western Kansas to be Quaternary in age and to be at least in part of eolian origin. None of the early workers in Kansas, however, recognized subdivisions within the body of the silt or understood the relationship of silt deposition in the Plains region to the cyclic episodes of continental glaciation to the east, north, and west.

In northeastern Kansas, loess deposits were noted by Swallow (1866, pp. 36–38) and Hawn (1866, pp. 99–101), and in 1917 Hinds and Greene (p. 8) discussed the possibility of correlation of deposits in this region with the Loveland and Peorian of Iowa.

Present classification of the western Kansas loesses has developed by subdivision of deposits included by Elias (1931) in the Sanborn formation. He considered these deposits to extend generally over northwestern Kansas and proposed the name "Sanborn" from a locality (pl. 1, *D*) in northwestern Cheyenne County, where they (Crete member with Sangamon soil and Peoria member) are well exposed along the south bluff of the Arikaree River Valley. Further, Elias recognized three phases of deposition which display general similarity of appearance within the body of massive silt: (1) eolian deposits on the divides; (2) colluvial or predominantly gravity deposits on the slopes (pl. 2, *E*); and (3) reworked fluvial deposits in some valley bottoms. He restricted the Sanborn formation to the eolian sediments of the divide areas. In this paper the term "loess" is applied only to these eolian deposits properly included within the Sanborn formation.

Since the work by Elias, studies in Nebraska (Lugn, 1935; Schultz and Stout, 1945, 1948; Condra, Reed, and Gordon, 1947) and Kansas have made possible the stratigraphic subdivision of the materials included within the Sanborn and the correlation of these units with described stratigraphic units in adjacent states. Subdivisions have been based on lithologic differences, regional unconformities, and distinctive faunal assemblages. The three silt members do

not possess lithologic distinctiveness except where they have been altered by former episodes of weathering, but the lowest member—the Crete sand and gravel (though gradational with the overlying Loveland silt)—is clearly distinguishable lithologically. The three silt members are separable both by regional unconformities evidenced by prominent buried soil profiles and by their contained faunas. A developed zonal profile in a succession of clastic sediments constitutes conclusive evidence of unconformity, inasmuch as the surface below which such a profile developed must have been exposed to weathering without the influence of excessive erosion or deposition. Buried soil profiles, therefore, are judged to be valuable criteria for stratigraphic subdivision, and such features are used to define the boundaries of the three silt members. Assemblages of fossil mollusks have been collected widely from these units (fig. 1), as well as from the underlying early Yarmouthian Sappa member of the Meade formation (Leonard, 1950) throughout the same region. These faunal assemblages of the Sanborn formation as now known from Kansas are shown in figure 2. They have proved to be of value not only for field identification of the several members but also for regional correlations. Fossil mammals have been collected from the Sanborn formation at several localities in Kansas (Hibbard, Frye, and Leonard, 1944), but their sparse occurrence makes them of small value for recognition of the several units in the field.

Current usage in Nebraska (Condra, Reed, and Gordon, 1947) assigns formational rank to each of the four members here assigned to the Sanborn formation. Even though the similarly named units in Nebraska are stratigraphically continuous with the units in Kansas, the assignment of member, rather than formational, rank in the Kansas region is judged to be preferable, because (1) in the course of field mapping on a scale of 1 inch = 1 mile it is generally impracticable or impossible to map the several loess units separately; (2) in some areas of very thin leached loess, definite identification of the several units is difficult or impossible; and (3) the grouping within one formational unit of all post-Yarmouthian Pleistocene sediments serves to emphasize the sharp difference between the geologic history of

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PLATE 1

A and *B*, Bignell loess, Brady soil, and Peoria loess at type locality of the Bignell loess and Brady soil south of Bignell, Nebraska, on south valley wall of Platte River Valley; Sec. 3, T. 12 N., R. 29 W., Lincoln County (Schultz and Stout, 1945, p. 241). Fauna locality 93. 1949.

C, Bignell silt member (5 feet) on Brady soil in the top of Peoria silt member (35 feet) on Sangamon soil developed in sandy silts of the Crete member, Sanborn formation. Crete member rests on Ogallala formation (Pliocene). Road cut along U.S. Highway 36, Sec. 2, T. 3 S., R. 33 W., Rawlins County, Kansas. Fauna locality 22. 1948.

D, Canyons in Pleistocene deposits in the type area of the Sanborn formation; SW. ¼ of SE. ¼, Sec. 17, T. 1 S., R. 41 W., Cheyenne County, Kansas. In adjacent tributary canyons 90 feet of Peoria silt member resting on Sangamon soil developed in upper part of Crete member was measured. Fauna locality 25. 1948.

E, Brady soil in road cut; SW. ¼ of SW. ¼, Sec. 24, T. 4 S., R. 19 W., Phillips County, Kansas. Photo by A. R. Leonard, 1947.

F, Loveland and Peoria loesses at the type locality of the Loveland, at Loveland, Iowa. Exposures are in the bluff of Missouri River Valley. The Sangamon soil appears as a dark band in the photograph approximately one-third of the way down from the top of the bluff. The Loveland rests on glacial till. Peoria is fossiliferous, and the Loveland is sparsely fossiliferous in lowermost part. Fauna locality 97. 1950.

A

B

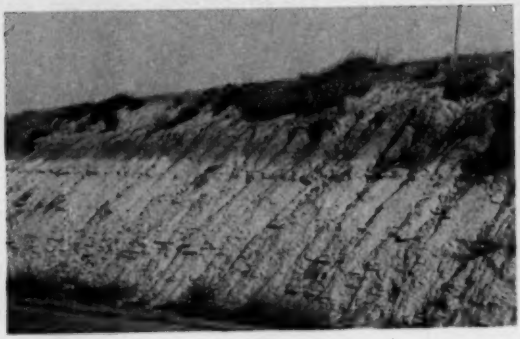
C

D

E

F

Late Pleistocene loesses



A

B

C

D

E

F

Late Pleistocene loesses

early and late Pleistocene time in Kansas.

ILLINOIAN-SANGAMONIAN

The Yarmouthian interglacial interval serves to divide the Pleistocene history of Kansas into two dissimilar parts. Early Pleistocene time was characterized by the invasion of the state by continental glaciers, the incision and filling of major valleys by glacially modified streams and their larger tributaries, and profound changes in drainage patterns. During late Pleistocene time, continental glaciers lacked 200 miles of reaching the northeastern corner of Kansas. There were few major alterations in the drainage patterns, cut-and-fill terraces devel-

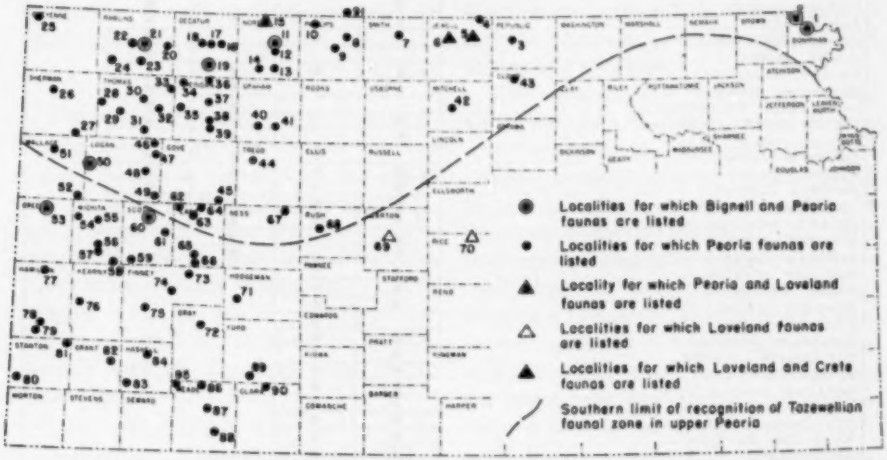

Localities for which Bignell and Peoria faunas are listed
Localities for which Peoria faunas are listed
Locality for which Peoria and Loveland faunas are listed
Localities for which Loveland faunas are listed
Localities for which Loveland and Crete faunas are listed
Southern limit of recognition of Tazewellian faunal zone in upper Peoria

FIG. 1.—Map of Kansas showing localities from which late Pleistocene molluscan faunas are listed. Locality numbers are those used in text and figs. 2 to 5; localities 92–94 are in Nebraska, 95 in Oklahoma, 96 in Texas, and 97 in Iowa.

PLATE 2

A, Peoria silt member (10 feet) on Sangamon soil developed in Loveland silt member (6 feet) of Sanborn formation, on Greenhorn limestone (Cretaceous). Exposure in cut on U.S. Highway 36, NW. ¼ of NW. ¼, Sec. 9, T. 3 S., R. 1 W., Republic County, Kansas. Photo by A. R. Leonard, 1948.

B, Vertical surface of Peoria silt member, northeastern Doniphan County, Kansas, showing color banding and soft vertically oriented limonite concretions. 1948.

C, Bignell silt member (35 feet) on Brady soil developed in upper part of Peoria silt member, Sanborn formation; exposed in State Highway Department excavation, Sec. 6, T. 2 S., R. 30 E., Doniphan County, Kansas. Fauna locality 1. 1948.

D, Colluvium on graded slope developed on Greenhorn limestone (Cretaceous); south side of Saline River Valley, south of Sylvan Grove, Lincoln County, Kansas. The uplands in this area are mantled thinly with Peoria loess, and graded or pediment-like slopes are veneered with a mantle of silt containing Greenhorn fragments in the lower part. As this mantle is in intermittent transit, it is not included within the Sanborn formation. 1948.

E, Thin upland mantle of Peoria silt member exposed in road cut on Kansas Highway 14, SW. ¼, Sec. 18, T. 10 S., R. 7 W., Lincoln County, Kansas. Sanborn formation rests on Greenhorn limestone (Cretaceous). 1949.

F, High Plains surface mantled with Peoria silt member; SE. cor. Sec. 35, T. 20 S., R. 35 W., Wichita County, Kansas. In an auger hole, 13.5 feet of Peoria silt was penetrated above Sangamon soil. Fossils were recovered from auger cuttings. Fauna locality 58. 1949.

Species	Sappa	Crete–Loveland	Peoria–Lower	Peoria–Upper	Bignell
Anguispira alternata					●
Stenotrema monodon alicia					●
Triodopsis albolabris					●
Vallonia pulchella	●				●
Gastrocopta armifera	●	●		●	●
Hendersonia occulta	●			●	●
Succinea grosvenori	●	●		●	●
Succinea ovalis	●			●	●
Hawaiia miniscula	●	●	●	●	●
Helicodiscus parallelus	●	●	●	●	●
Gastrocopta holzingeri	●	●		●	
Columella alticola				●	
Striatura milium				●	
Vertigo coloradensis				●	
Vertigo modesta	●			●	
Carychium exiguum	●	●		●	
Cionella lubrica	●			●	
Pupilla blandi	●	●	●	●	
Discus cronkhitei	●	●		●	
Discus shimeki				●	
Vertigo tridentata	●	●	●	●	
Euconulus fulvus			●	●	
Helicodiscus singleyanus			●	●	
Vertigo gouldi paradoxa	●		●	●	
Vertigo milium	●		●	●	
Deroceras laeve		●	●	●	
Pupilla muscorum	●	●	●	●	
Retinella electrina	●	●	●	●	
Zonitoides arboreus	●	●	●	●	
Vallonia gracilicosta	●	●	●	●	
Succinea avara	●		●		
Lymnaea parva	●		●		
Helisoma trivolvis	●	●			
Physa anatina	●	●			
Pupoides albilabris	●	●			
Strobilops labyrinthica		●			
Carychium pereziguum	●	●			
Gastrocopta cristata	●	●			
Gyraulus similaris	●	●			
Helisoma antrosa	●	●			
Sphaerium solidulum	●	●			
Gastrocopta tappaniana	●	●			
Strobilops sparsicosta	●	●			
Vertigo ovata	●	●			
Ferrissia parallela	●				
Gastrocopta contracta	●				
Gastrocopta falcis	●				
Gastrocopta procarmifera	●				
Gastrocopta procera	●				
Amnicola limosa parva	●				
Gastrocopta tridentata	●				
Gyraulus labiatus	●				
Gyraulus pattersoni	●				
Helisoma wisconsinensis	●				
Lymnaea bulimoides	●				
Lymnaea caperata	●				
Lymnaea palustris	●				
Lymnaea reflexa	●				
Polygyra texasiana	●				
Menetus pearlettei	●				
Oxyloma navarrei	●				
Physa elliptica	●				
Pisidium compressum	●				
Planorbula nebraskensis	●				
Planorbula vulcanata vulcanata	●				
Planorbula vulcanata occidentalis	●				
Pomatiopsis cincinnatiensis	●				
Promenetus umbilicatellus	●				
Pupilla muscorum sinistra	●				
Stenotrema monodon monodon	●				
Aplexa hypnorum	●				
Valvata lewisi	●				
Valvata tricarinata	●				
Deroceras denigma	●				

FIG. 2.—Check list of fossil mollusks from the Meade and Sanborn formations in Kansas

oped along existing valleys, large tracts of sand dunes were formed, and, of most widespread significance, upland and high terrace levels were blanketed with eolian silts. The Yarmouthian interglacial interval was a relatively long time of essential equilibrium, typified by adjustments of drainage to the dislocations and piracies initiated during Kansan time and by soil formation. The recognizable loesses of Kansas are all post-Yarmouthian in age.

CRETE MEMBER

The name "Crete formation" was assigned by Condra, Reed, and Gordon (1947, p. 24) to deposits of sand and gravel resting unconformably on the Upland (Sappa) formation at many localities in Nebraska. They considered the Crete to be Illinoian in age and to grade upward into the "valley phase" of the Loveland, which is stratigraphically continuous with the Loveland loess of the upland areas.

The Crete member is known to occur prominently in Kansas only in the north-central area (Frye and Fent, 1947; Frye and A. R. Leonard, 1949), although scattered deposits in southeastern, central, and southwestern Kansas are judged to be assignable to the member. In north-central Kansas, the Crete sands and gravels occur in a terrace position along several hundred miles of valley, particularly the valleys of Sappa and Prairie Dog creeks and Solomon, Saline, and Smoky Hill rivers. In the northern tier of counties these deposits do not form a distinct topographic terrace, because they are partly dissected and are masked by younger loess deposits that extend from the former terrace surface up the valley-side slopes to the upland level. The relation of the Crete sand and gravel along these valleys to the younger low terraces is clearly shown by test drilling (Frye and A. R. Leonard, 1949). The lithology of the Crete sands and gravels generally reflects a local source. In the north-central valleys the Ogallala formation and locally earlier Pleistocene deposits were extensive in the drainage basins, and rocks typical of the Rocky Mountain region are prevalent. In the Saline Valley and some of the minor valleys of the Cretaceous outcrop belt, the gravels are largely composed of Cretaceous sandstones and limestones.

The stratigraphic placement of the Crete member of north-central Kansas is well established. The next older unit—the Sappa member of the Meade formation—has been correlated over the mid-continent region by the distinctiveness of its contained Pearlette volcanic ash lentil (Swineford and Frye, 1946; Frye, Swineford, and Leonard, 1948), and by a molluscan fauna (fig. 2) consisting of 65 species (Leonard, 1950). Both the ash and the fauna have been placed stratigraphically in the glacial sequence of the Missouri Valley region. The Meade formation is shown to be latest Kansan or early Yarmouthian in age. The Crete member is determined to be appreciably younger than the Meade, as indicated by physiographic relations and by its stratigraphic position above a soil profile on the Sappa silt. It is shown to be appreciably younger than Iowan by relation of the widely developed soil profile on the overlying Loveland to deposits of Iowan age described here. The molluscan fauna of the Crete and overlying Loveland is reported in figure 3. The association of forms typical of the older Sappa fauna (which make their last appearance here) with other species typical of the overlying Peoria fauna (which make their first appearance in this assemblage) serve to give the Crete-Loveland fauna distinctiveness for

correlation of the unit. However, this fauna has not as yet been obtained in close association with Illinois glacial till, and the dating of the Crete member as Illinoian rests largely on its stratigraphic relationships.

Elsewhere in Kansas the Crete member has been studied in less detail and is recognizable only at scattered localities. Minor terraces along some valleys of east-central Kansas may be corollary to the Crete. In southwestern Kansas some minor sand and gravel deposits overlying the Meade formation and included within units variously called "upper Meade" (Frye and Hibbard, 1941, p. 415) and "lower Kingsdown" (Hibbard, 1944; Frye, 1945), seem quite certainly to be classed properly as Crete. The geographic extent of the Crete of southwestern Kansas is small but has importance because of its contained vertebrate faunas.

LOVELAND MEMBER

The Loveland silt member of the Sanborn formation consists of water-depos-

Species	Locality Number					
	5	6	15	69	70	97
Anodonta sp.		●				
Carychium exiguum	●					
Carychium perexiguum				●	●	
Deroceras laeve					●	
Discus cronkhitei						●
Gastrocopta armifera	●					●
Gastrocopta holzingeri						●
Gastrocopta tappaniana		●				
Gyraulus similaris	●			●		
Hawaiia miniscula	●					
Helicodiscus parallelus	●	●				
Helicodiscus singleyanus	●					
Helisoma antrosa	●				●	
Helisoma trivolvis lentum	●	●				
Physa anatina	●				●	
Pupilla blandi		●				
Pupilla muscorum			●			
Pupoides albilabris	●	●				
Retinella electrina				●		
Sphaerium solidulum	●	●		●		
Strobilops sparsicosta					●	
Strobilops sp.						●
Succinea grosvenori	●	●	●	●		●
Vallonia gracilicosta	●	●	●			●
Vallonia pulchella	●					
Vertigo ovata						●
Vertigo tridentata			●			
Total number species	14	9	4	5	5	7

FIG. 3.—Fossil mollusks from the Loveland loess at Loveland and near Little Sioux, Iowa, and from the Crete and Loveland members at localities 5, 6, 15, 69, and 70 in Kansas.

ited silts which overlie gradationally the Crete member in some valley areas, but to a much greater extent it comprises widely distributed eolian silts on upland areas. The term "Loveland silt member" is applied in Kansas both to the water-laid silts in valley situations and to the loess of the uplands. The two facies of the silt unit have been determined to be stratigraphically continuous along some valley-side slopes, occupy the same stratigraphic position, are the same age, and can be mapped together conveniently.

The Loveland loess was named by Shimek in 1910 (p. 405) from exposures at the town of Loveland, Iowa (pl. 1, *F*), along the east bluff of the Missouri Valley. At the type section the Loveland is approximately 30 feet thick, rests on glacial till, and is overlain by typical fossiliferous Peoria loess. The upper 22–24 feet are leached of calcium carbonate and are included within the Sangamon (Loveland) soil profile. The part of the material included in the profile has a pronounced reddish tint, but the unleached basal part has a yellowish-tan color, common to many exposures of Peoria. Particle-size distribution of unleached silt from the lower part of the Loveland type section is similar to the size distribution of Loveland loess in Kansas and Peoria loess from many localities (personal communication from Ada Swineford). A meager but distinctive assemblage of fossil mollusks was obtained from the basal part of the Loveland type section (fig. 3). Deposits correlated with the Loveland loess have been traced widely throughout Nebraska (Condra, Reed, and Gordon, 1947), and the unit called "Loveland" in Nebraska is continuous with the deposits assigned to the Loveland member in Kansas. The lithology, stratigraphic sequence, characteristic morphology of the buried Sangamon soil, and data from the sparse fauna of the Loveland and the abundant widespread faunas of the underlying Sappa and overlying Peoria serve to substantiate correlation of the Kansas Loveland with the Loveland of the type locality.

The Loveland, although it is the earliest important Pleistocene loess of the Great Plains region, is relatively thin and discontinuous in comparison to the overlying Peoria. In Kansas the Loveland loess is well developed in the northern half and central part of the state. It is commonly 5–15 feet thick along the Missouri Valley bluffs of northeastern Kansas, and westward it is thin to indistinguishable in much of the area east and southeast of northern Washington County. Westward from Republican Valley for 200 miles, the Loveland silt is a thin, but distinctive and persistent, unit in northern Kansas (Hibbard, Frye, and Leonard, 1944; Frye and Fent, 1947; Frye *et al.*, 1949; Frye and A. R. Leonard, 1949). In northwestern Kansas thin Loveland silt occurs only at a few places, although the Crete member is relatively widespread and the Sangamon (Loveland) soil is distinctively developed on Crete and older deposits.

The Loveland silt seems to attain its maximum thickness in Rice and McPherson counties in central Kansas, where thicknesses of 80 and 90 feet have been reported as occurring (Frye and Fent, 1947, p. 45; Fent, 1950*b*) in earlier Pleistocene valleys, now abandoned. Recent work, however, has shown that in some places in central Kansas an earlier miscorrelation of upper Sappa silts in the subsurface as Loveland, where neither Crete member nor Sappa (Yarmouth) soil intervened, has resulted in the reporting of excessive thicknesses of Loveland. Sappa silts are water-deposited, interstratified with fine sand, and contain the Pearlette volcanic ash lentil, which makes possible their correlation. The

Loveland silt in McPherson County may not exceed 50 feet in thickness.

A particularly significant loess exposure occurs in northern Rice County (Frye and Fent, 1947, p. 50; Frye *et al.*, 1949, loc. 21; loc. 70 in this paper). Here the three loess units of the Sanborn formation occur in an upland situation, resting on the weathered surface of Cretaceous shale. The Peoria and Bignell members are leached throughout, but a thin calcareous zone in the basal part of the Loveland contains a relatively rich molluscan fauna.

In southwestern Kansas eolian silts equivalent to the Loveland member are rare to nonexistent. Dune sands have been observed resting on Sappa silts, the sands, in turn, terminating upward in a buried soil which is overlain by one or more units of younger dune sand. In some places in the southwest, eolian sands may be stratigraphic equivalents of the Loveland member, but the sands which give topographic expression to the existing tracts are much younger.

SANGAMON (LOVELAND) SOIL

The feature most distinctive of the Loveland, both in northern Kansas and throughout the central Missouri Valley region, is the prominent and distinctive soil profile commonly developed in the upper part of the silt during Sangamonian time. This soil is perhaps the most widespread continuously traceable datum in the Pleistocene succession of the Midwest. In the Loveland type section the soil is well developed (pl. 1, *F*), with a profile approximately 25 feet thick. The zone of clay accumulation is 2–3 feet thick, shows well-developed structure, and is light red-brown in color; the upper 22–24 feet of the soil are leached of calcium carbonate, and a zone of caliche accumulation occurs at the base of the leached zone.

The Sangamon (Loveland) soil has been traced into areas where it is overlain by Iowa glacial till, showing that the time of soil development was terminated by the advance of Iowan ice. Beyond the limits of Wisconsinan continental ice sheets, the Loveland soil is commonly overlain by Wisconsinan loess, classed in Kansas as the Peoria silt member. In Kansas the loess overlying the Sangamon soil is generally transitional in the lower part. That is, after the Sangamon soil profile had reached its maximum development, increments of silt were added to the top of the "A horizon" at a rate allowing their incorporation into the soil profile. These added increments of silt were deposited progressively faster and therefore were leached less, so that the partly weathered silts resting on the mature profile grade upward into fresh-appearing calcareous, fossiliferous loess. At a few places incipient soils have been observed in this basal zone. For the sake of uniformity, the top of the Sangamon (Loveland) soil, as used here, is the top of the initial "A horizon" that marks the top of the mature profile and below the superimposed immature "A horizon," here classed as the basal part of the overlying Peoria member. In this connection it should be emphasized that the *Citellus* zone—a faunal zone extensively recognized in Nebraska—transgresses this member boundary (Condra, Reed, and Gordon, 1947, p. 30), although it is clear that the mammals whose remains characterize the faunal zone lived after the Sangamon (Loveland) soil was formed. This seeming anomaly arises from the fact that the fauna of the *Citellus* zone is typified by burrowing mammals, some individuals of which burrowed deeply into the soil profile—or, indeed, entirely through it—whereas other mammals included within the fauna lived on the soil surface and their remains are found in

the basal part of the Peoria member. Filled rodent burrows containing fossil remains are found within the Sangamon profile.

The characteristics of profile morphology displayed by the Sangamon have a wide range, even when considered from localities of similar situation (Frye, 1949). Along the Missouri River bluffs northward from Kansas City and in Marshall, Washington, and eastern Republic counties, the Sangamon soil profile contains the entire thickness of the member. Sediments correlatable with the Loveland member have not been recognized in the central part of the Kansas glaciated area, and it is possible that at some places the profile in the upper part of the Kansas till, buried by Peoria loess, may be the product of weathering during both Yarmouthian and Sangamonian time. The Sangamon soil has its most striking appearance in outcrop sections in the area extending from Republic County westward to Decatur County. In this area the upper part of the profile grades from gray-black through reddish brown to pink-buff (pl. 2, *A*), the depth of leaching is commonly 3–5 feet, and the zone of caliche accumulation is 1–3 feet thick. Although thin Loveland silt contains the profile in much of this area, the soil has been observed in stratigraphic sequence below Peoria loess but developed on chalky shales of Cretaceous age, Crete sands and gravels, and colluvial slope deposits.

Westward from Decatur County the Sangamon soil, although lighter in color and incompletely leached, continues to be a distinct stratigraphic datum. The profile is commonly developed on sandy silts or sands of the Crete member in extreme northwestern Kansas, as is the case at the type section of the Sanborn formation (pl. 1, *D*).

In central Kansas, Sangamon (Loveland) soil is widely developed on the Loveland silt member (Frye and Fent, 1947), its color is more strongly red than farther north, and at two localities (69 and 70) fossil mollusks have been obtained from the unleached silts at the base of the profile. A buried soil occurs widely at shallow depths throughout the area included within the great bend of the Arkansas River and, although scantily exposed, has been penetrated by auger and drill holes. This soil was earlier identified in the field as Sangamon (Loveland). The relation of the containing silts to Pearlette volcanic ash lentils and faunas and the more strongly developed profile itself suggest that it is developed on the Sappa member of the Meade formation. This soil underlies much of the area of the Great Bend sand-dune tract.

The profile morphologies in southwestern Kansas present a marked contrast to those of northwestern and north-central Kansas. Eolian silts of the Loveland member have not been identified in southwestern Kansas, but Peoria loess containing a diagnostic fauna has been traced extensively over the area (fig. 1). Peoria loess at many places rests on a deep profile characterized by strong red color and a zone of nodular caliche as much as 4 or 5 feet in thickness. At several places the sandy silts in which this deep profile is developed can be correlated by volcanic ash or fauna as the Sappa member of the Meade formation. Buried soil profiles, where observed on silty sands of the Crete member, are less strongly developed. These facts lead to the judgment that at some places in southwestern Kansas the Sappa member was little eroded throughout Yarmouthian, Illinoian, and Sangamonian time and that this span of time resulted in the formation of a single profile. Because the climax of formation of this soil came during Sangamonian time and because it is

immediately overlain by Peoria loess, there is little choice but to class it as Sangamon (Loveland) soil developed in the Sappa member of the Meade formation.

EARLY WISCONSINAN

After the interval of conditions favorable for soil formation (Sangamonian interglacial interval), the first effect of the oncoming early Wisconsinan glaciation in the Kansas region was change in the regimen of through-flowing streams. Some valley floors that had persisted during much of Sangamonian time were dissected, and in places the channels were cut deeply into bedrock. Later these same valleys were alluviated. The channels of some tributary streams became incised, but regionally there was little erosion of intermediate and high-level surfaces. Rather, eolian silts were deposited on the relatively uneroded surface of the soil developed during Sangamonian time.

Deposits of eolian silt furnish the most widespread record of early Wisconsinan time. These silts are determined to have been derived from the floodplains of the major outwash-carrying valleys of the region (Swineford and Frye, 1951). The rate of loess deposition, which was largely controlled by availability of silts along these valleys, varied. Generally it started slowly, accelerated to a relatively rapid rate, slowed rapidly, and then ceased. Mid-Wisconsinan time in Kansas was again an interval of soil formation, followed during late Wisconsinan by a period of less extensive loess deposition. In many northern Kansas valleys, prominent terraces, which commonly include a soil profile, show the same succession of events.

Continental glaciers of Wisconsinan age invaded the Missouri Valley in northern Iowa and Nebraska and northward, and mountain glaciers were active in the central Rocky Mountain region. The Wisconsinan history of Kansas is judged to have been controlled largely by the activity of these ice masses in the headwaters of major drainage systems rather than by profound climatic changes within the state.

In the Kansas region the Wisconsinan loesses were derived principally from outwash-carrying streams which flowed in a generally easterly direction and derived their sediment load from the west. It is not known with certainty that the several episodes of Wisconsinan glaciation in the Rocky Mountain region were synchronous with those which invaded the north-central interior. The basal zone of the Peoria possesses some characters indicative of an early Iowan history in the mountains that was slightly different from the early Iowan history of the central interior. Loess called "Farmdale" in Illinois (Leighton and Willman, 1949) and loess in Iowa at the base of the Peoria which is judged to be equivalent to the Farmdale show evidence of a distinct slowing in the rate of loess accumulation or possibly a pause in deposition before the main body of Peoria was laid down. However, there is almost complete gradation from the leached basal part of the Peoria upward into calcareous loess, and along the Platte Valley of western Nebraska several incipient soils occur in the basal Peoria (Schultz and Stout, 1945). Whereas the Farmdale and its equivalents may imply a pulsation in the advancing Iowan ice of the interior, the loess of the plains furnishes evidence of more continuous glacial advance. Since the basal zone in the plains does not contain a molluscan fauna, its contemporaneity with the Farmdale of Illinois is indicated only by its stratigraphic posi-

tion on the Sangamon (Loveland) soil and below fossiliferous loess of Iowan age.

PEORIA MEMBER

Leverett in 1898 (1898*a*, p. 175) applied the name "Iowan loess" to deposits adjacent to the margin of Iowan till. In the same year (Leverett, 1898*b*, p. 248), because he discovered a weathered zone at the top of this loess, he applied the name "Peorian" to a presumed post-Iowan interglacial interval. Alden and Leighton (1917) showed that the Iowan loess was actually post-Iowan (Peorian) in age, and usage gradually changed the name of the loess from Iowan to Peorian. "Peorian" has been used throughout Iowa to apply to loess of post-Iowan, pre-Mankatoan age (Kay, 1931; Kay and Graham, 1943) and to equivalent deposits across Nebraska (Lugn, 1935; Schultz and Stout, 1945; Condra, Reed, and Gordon, 1947). Loess deposits within the border of the Tazewell till in Illinois are interstratified with tills and are called "Farmdale," "Iowan," and "Tazewell" (Leighton and Willman, 1949); but, beyond the margin of that till, Iowan and younger loesses are generally not differentiated and are collectively referred to as "Peorian loess" (Smith, 1942).

As now used in Kansas, the Peoria silt member of the Sanborn formation includes loess and locally water-laid silt that is stratigraphically continuous with it (Frye and Fent, 1947; Frye, Swineford, and Leonard, 1948; Frye and A. B. Leonard, 1949). Stratigraphic and faunal data (Leonard, 1951) show the Peoria of Kansas to be equivalent to the Peorian loess of Nebraska (Condra, Reed, and Gordon, 1947) and Iowa (Kay and Graham, 1943), and to the loesses classed as Farmdale, Iowan, and Tazewell in central Illinois. The term "Peoria" is deemed appropriate for this loess in Kansas because of its equivalency to loess beyond the margin of the Tazewell till in the vicinity of Peoria, Illinois, where the loess is known to contain equivalents to the Iowan and Tazewell loesses immediately adjacent to the east.

The term "Peoria loess" is used in this paper to designate the eolian silts on upland divides and high terrace levels, where the silts have not been moved appreciably by agencies other than the wind. The term "Peoria silt" includes, in addition to the loess, stratigraphic equivalents to it that were water-deposited or modified by colluvial processes on slopes. Colluvial slope veneers, other than those that represent downward creep of loess from higher levels, may have little relation to the Peoria loess (even though of approximately the same age) and are not included within the Peoria silt member (pl. 2, *D*).

The Peoria loess throughout Kansas is strikingly uniform in composition (Swineford and Frye, 1951), general appearance and fauna. Figure 4 shows the significant elements of the molluscan fauna of the Peoria from 92 localities: 86 in Kansas (fig. 1), 3 in Nebraska, and 1 each in Texas, Oklahoma, and Iowa. The loess in the central Great Plains is thickest adjacent to the principal source valleys (Swineford and Frye, 1951)—namely, the Platte in Nebraska, the Missouri of northeastern Kansas, and, to a less extent, the Arikaree and Republican rivers.

The Peoria is shown by test drilling in northeastern Kansas (Frye and Walters, 1950) to attain a maximum thickness of more than 100 feet, but it thins markedly in a relatively short distance away from the valley toward the west, southwest, and south. Some Peoria loess may be present as far south in eastern Kansas as the Oklahoma state line, but south of the

| Species \ Location | SE sec. 6, T. 2 S., R. 20 E., Doniphan Co., Kan. | NE sec. 8, T. 1 S., R. 19 E., Doniphan Co., Kan. | NW sec. 5, T. 3 S., R. 4 W., Republic Co., Kan. | SE sec. 4, T. 1 S., R. 7 W., Jewell Co., Kansas | SE sec. 28, T. 1 S., R. 13 W., Smith Co., Kansas | SE sec. 5, T. 2 S., R. 17 W., Phillips Co., Kansas | NW sec. 23, T. 2 S., R. 18 W., Phillips Co., Kansas | NE sec. 25, T. 1 S., R. 20 W., Phillips Co., Kansas | NW sec. 26, T. 2 S., R. 23 W., Norton Co., Kansas | NW sec. 11, T. 3 S., R. 23 W., Norton Co., Kansas | NE sec. 10, T. 4 S., R. 23 W., Norton Co., Kansas | NE sec. 7, T. 5 S., R. 24 W., Norton Co., Kansas | SE sec. 13, T. 1 S., R. 24 W., Norton Co., Kansas | NE sec. 3, T. 3 S., R. 27 W., Decatur Co., Kan. | NE sec. 6, T. 3 S., R. 27 W., Decatur Co., Kan. | NE sec. 1, T. 3 S., R. 28 W., Decatur Co., Kan. | SW sec. 23, T. 4 S., R. 27 W., Decatur Co., Kan. | NE sec. 2, T. 3 S., R. 31 W., Rawlins Co., Kan. | SW sec. 2, T. 3 S., R. 33 W., Rawlins Co., Kan. | NE sec. 9, T. 3 S., R. 33 W., Rawlins Co., Kan. | NW sec. 20, T. 4 S., R. 33 W., Rawlins Co., Kan. | NE sec. 29, T. 3 S., R. 36 W., Rawlins Co., Kan. | SE sec. 17, T. 1 S., R. 41 W., Cheyenne Co., Kan. | NW sec. 32, T. 7 S., R. 39 W., Sherman Co., Kan. | SW sec. 26, T. 10 S., R. 38 W., Sherman Co., Kan. | SW sec. 7, T. 8 S., R. 36 W., Thomas Co., Kansas | NW sec. 8, T. 8 S., R. 35 W., Thomas Co., Kansas | NW sec. 31, T. 7 S., R. 33 W., Thomas Co., Kansas | SE sec. 27, T. 10 S., R. 33 W., Thomas Co., Kansas | NE sec. 17, T. 8 S., R. 32 W., Thomas Co., Kansas | SE sec. 21, T. 7 S., R. 31 W., Thomas Co., Kansas | NW sec. 15, T. 6 S., R. 30 W., Sheridan Co., Kan. | NW sec. 31, T. 8 S., R. 30 W., Sheridan Co., Kan. | SW sec. 15, T. 6 S., R. 28 W., Sheridan Co., Kan. | SE sec. 28, T. 7 S., R. 28 W., Sheridan Co., Kan. | NW sec. 34, T. 8 S., R. 28 W., Sheridan Co., Kan. | SW sec. 10, T. 9 S., R. 28 W., Sheridan Co., Kan. | NE sec. 28, T. 9 S., R. 22 W., Graham Co., Kan. | SW sec. 23, T. 9 S., R. 21 W., Graham Co., Kan. | NW sec. 28, T. 8 S., R. 8 W., Mitchell Co., Kan. | NW sec. 31, T. 5 S., R. 3 W., Cloud Co., Kansas | SW sec. 5, T. 12 S., R. 24 W., Trego Co., Kansas |
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SW sec. 5, T.12S.,R.24W., Trego Co.,Kansas
NE sec. 35, T.15S.,R.27W, Gove Co., Kansas
SE sec. 27, T.11S.,R.32W, Logan Co.,Kansas
NW sec. 13, T.12S.,R.32W.,Logan Co.,Kansas
SE sec. 3, T.14S.,R.32W., Logan Co.,Kansas
NW sec.27, T.15S.,R.32W., Logan Co.,Kansas
NE sec. 32,T12S., R.37W, Logan Co.,Kansas
SW sec. 35, T.11S., R.40W., Wallace Co.,Kan.
NW sec. 26, T.15S., R.38W., Wallace Co.,Kan.
NW sec. 9, T.16S.,R.40W., Greeley Co.,Kan.
SE sec. 32, T.16S.,R.38W., Wichita Co.,Kan.
SW sec. 12, T.17S., R.37W., Wichita Co., Kan.
SE sec. 1 , T.19S.,R.37W., Wichita Co.,Kan.
NE sec.12, T.19S.,R.37W., Wichita Co.,Kan.
SE sec.35, T.20S.,R.35W., Wichita Co.,Kan.
SW sec. 31, T. 20S.,R.34W.,Scott Co.,Kansas
SE sec. 6, T. 17S.,R.32W., Scott Co.,Kansas
SE sec.17, T.18S., R.31W., Scott Co.,Kansas
E½ sec. 15, T.16S., R.30W., Lane Co.,Kansas
NE sec.15, T.16S., R.29W., Lane Co.,Kansas
NW sec.3, T.16S., R.29W., Lane Co.,Kansas
NW sec.7, T.20S., R.28W., Lane Co.,Kansas
SE sec. 29, T.20S., R.28W., Lane Co.,Kansas
NW sec.10, T.16S., R.22W., Ness Co., Kansas
SW sec. 28, T.17S., R.19W., Rush Co., Kansas
NW sec. 26,T.23S., R.25W.,Hodgeman Co.,Kan.
NE sec. 1 , T. 25S., R. 29W., Gray Co.,Kansas
SW sec. 24,T.21S., R.29W., Finney Co.,Kan.
NE sec. 34,T.22S., R.30W.,Finney Co., Kan.
SW sec. 9, T. 24S., R.32W., Finney Co.,Kan.
NW sec. 31, T. 23S., R.38W., Kearny Co.,Kan.
NW sec. 19, T.21S. , R.40W., Hamilton Co., Kan.
SE sec. 12, T.25S., R.41W., Hamilton Co.,Kan.
SW sec. 36,T. 25S.,R.41W., Hamilton Co.,Kan.
SE sec. 25 , T.29S.,R.43W., Stanton Co.,Kan.
SW sec. 11, T.27S.,R.39W., Stanton Co., Kan.
SE sec. 28, T.27S., R.35W., Grant Co.,Kansas
NE sec. 22, T.30S.,R.34W., Haskell Co.,Kan.
NW sec. 3, T.28S., R.32W., Haskell Co., Kan.
NW sec. 6, T.30S.,R.30W., Meade Co.,Kansas
NE sec. 11, T.30S.,R.28W., Meade Co.,Kansas
SW sec.30, T.31S., R.28W., Meade Co.,Kansas
NE sec. 8, T. 33S., R.27W., Meade Co.,Kansas
NW sec. 16, T.29S.,R. 24W., Ford Co.,Kansas
NW sec. 6, T. 30S., R.23W.,Clark Co.,Kansas
NE sec. 31, T.1N., R.17W., Harlan Co.,Nebr.
Sec. 9 , T. 7 N., R. 29 W., Frontier Co.,Nebr.
E½ sec.3, T.12N., R. 29W., Lincoln Co., Nebr.
SE sec.7, T.10N., R.26W., Lincoln Co., Nebr.
1.8 miles N. Talogo, Dewey Co., Oklahoma
13.8 miles S. Stratford, Sherman Co.,Texas
Sec. 3, T.77S.,R.44W.,Pottawattamie Co.,Iowa
3 44 45 46 47 48 49 50 51 52 53 54 55 56 57 58 59 60 61 62 63 64 65 66 67 68 71 72 73 74 75 76 77 78 79 80 81 82 83 84 85 86 87 88 89 90 91 92 93 94 95 96 97
0 6 4 4 4 7 7 6 6 3 5 3 5 3 5 4 6 5 8 4 7 4 4 7 4 10 4 4 4 6 6 5 6 3 5 5 5 4 4 5 4 4 4 4 8 6 4 14 8 11 12 4 4 12

ocalities in Iowa, Kansas, Nebraska, Oklahoma, and Texas

glacial border it is generally present in thin discontinuous patches, is entirely leached, included within the profile of the topsoil, and devoid of fauna. In eastern Kansas, only in Doniphan County (pl. 2, *B* and *C*) does Peoria loess importantly influence the topography. Here, particularly in the vicinity of White Cloud (fauna loc. 2) and for 15 miles toward the south and east, the topography consists essentially of loess hills, but they do not show a distinct preferred orientation.

The upland till plains west of the Missouri Valley near the Nebraska state line are thinly veneered with Peoria. The loess again has prominent topographic expression along the Republican Valley in Republic County. As much as 40 feet of Peoria have been measured overlying Sangamon (Loveland) soil in the east bluff of the Republican Valley (loc. 3). West of the valley, this loess is only a few feet thick where it underlies the surface of an extensive plain coextensive with the outcrop belt of the Carlile shale.

Peoria loess thinly mantles broad flat upland divide areas generally underlain by Greenhorn limestone and the Fairport member of the Carlile shale (pl. 2, *E*) southwestward from Republic County across Mitchell and Lincoln counties The Peoria commonly does not exceed 5 feet in thickness on upland areas adjacent to the Smoky Hill Valley in Ellsworth, Russell, and Ellis counties; but farther south, in McPherson, Rice (Fent, 1950*b*), and Barton (Latta, 1951) counties, thicknesses of 10–20 feet have been observed.

West of the Republican River Valley in northern Kansas, a sharp topographic discordance occurs at the Fort Hays limestone escarpment. The Peoria occurs with almost uniform thickness and character above the Sangamon (Loveland) soil both on the Carlile shale plain and as much as 200 feet higher on the crest of the adjacent escarpment in Jewell, Smith, and Osborne counties.

Peoria loess is uniform in appearance and is highly fossiliferous throughout the area extending westward from Jewell County to Cheyenne County. Concentrations as high as 5,000 snail shells per cubic foot are common from localities 8 through 24 and 29 through 40. The Peoria attains thicknesses of 30–40 feet (pl. 1, *C*), and stratigraphic zonation (Leonard, 1951) of the member by molluscan faunas is possible at most exposures within this area. In the tier of counties adjacent to the Colorado state line (pl. 1, *D*) the concentration of fossil shells is sharply diminished, but they are in sufficient numbers to be diagnostic of the member (locs. 25, 26, 27, and 51).

The flat upland surface of the High Plains is relatively undissected (pl. 2, *F*) south of Smoky Hill Valley, in southern Wallace and Logan counties, throughout much of Greeley, Wichita, Scott, and Lane counties, and in northern Hamilton, Kearny, and Finney counties. Peoria loess, locally overlain by thin Bignell loess, underlies nearly all the High Plains surface throughout this region. The Peoria is commonly less than 10 feet thick, but the lack of exposures made extensive augering necessary in order to obtain faunas and stratigraphic data (locs. 54–59, 65, 66, 72, and 74–77). Within this region (fig. 1), the upper Peoria faunal zone of Tazewellian age has its southernmost extent (Leonard, 1951).

Immediately adjacent to the Arkansas River Valley, the Peoria, which caps terrace surfaces, shows some admixture of eolian sand (Swineford and Frye, 1951) but contains the typical fauna of the uplands (loc. 75). South of the Arkansas Valley, the Peoria loess shows a continu-

ation of the progressive thinning from north to south and, outside the tracts of sand dunes, constitutes the surficial material of much of the flat upland surface (locs. 78–87).

The Peoria silt member is predominantly loess in southwestern Kansas, but at a few places water-laid silts are included within the member. Outstanding among such localities is 88 in Meade County, where silts containing a distinctive Peoria molluscan fauna form the upper part of the dissected fill in a Pleistocene sink. Locality 89 in Ford County and near-by locality 90 in Clark County illustrate fauna containing aquatic elements (89) in silts that grade upward into loess and the more typical terrestrial assemblage (90) of the upland loess deposits.

Southward from southwestern Kansas, Peoria fauna was obtained from localities in Oklahoma (95) and Texas (96). Peoria loess caps the divide area between North and South Canadian River valleys in western Oklahoma and rests on Ogallala formation and Permian redbeds. The loess seems, on field inspection, to be much coarser than the upland material of southwestern Kansas and may be derived from a source along the valleys of the Canadian rivers. Thin Peoria loess is widely distributed in the northern part of the Texas panhandle as the surficial material, and in some places it forms low rounded swells in the surface, which rise a few feet above the general High Plains level. The fauna reported for locality 96 was obtained from a shallow cut in such a swell.

The topography in much of western Kansas is influenced by the Peoria loess, although prominent loess hills are not in evidence. A short-grass flora is judged to have prevailed in the High Plains region at the time of Peoria deposition, and drifting and reworking of the silts probably occurred. Aligned shallow depressions may locally mark the position of former shallow drainageways in the High Plains upland surface. Some major valleys contain thick Peoria (locs. 25 and 51) where adjacent deposits on higher points on the Ogallala surface are much thinner. In Cheyenne County elongate loess ridges occur approximately aligned with the inferred wind direction.

The Peoria cover is thinner farther south in the High Plains than in the northwestern area, and drift into east-west-trending valleys has a different expression. Localities 56 and 57 in Wichita County show differential drift into the north side of such a valley. At locality 56 the loess is thin and at the upland level; south of that locality the surface slopes down to the valley flat and exposes Ogallala in the shoulder of the valley wall. At the midpoint of the convex valley-side slope, the loess is thicker than under the upland surface and more abundantly fossiliferous (57), but along the south valley wall Ogallala formation is exposed in steep bluffs.

Peoria loess surficially seems to grade into dune sands at some places in southwestern Kansas, and at other places fossiliferous Peoria and its contained Brady soil are overlain by eolian sands (loc. 80).

BRADY SOIL

A soil profile universally occurs in the top of the Peoria loess. It is not given a stratigraphic name where the Peoria is at the surface and where this soil is the topsoil developed throughout post-Tazewellian time. Where the Bignell silt member overlies the Peoria, the soil at the top (pl. 1, *A*, *B*, *C*, and *E;* pl. 2, *C*) was formed in post-Tazewellian time but was terminated by the initial Bignell deposition. This fossil soil was named "Brady"

from a locality (pl. 1, *A* and *B*) in western Nebraska (Schultz and Stout, 1948, p. 570), and that name has been used generally throughout northern Kansas (Frye and Fent, 1947; Frye and A. B. Leonard, 1949; Frye *et al.*, 1949).

Brady soil has been examined along the Missouri bluffs of northeastern Kansas (loc. 1), adjacent to Republican River Valley and westward, in Rice County (loc. 70), and at many places in the northwestern quarter of the state (locs. 11, 19, 21, 50, 53, and 60). At most places where Brady soil has been studied, the topography on the soil surface and the morphology of the profile suggest that the soil formed under conditions of poor drainage, perhaps comparable to many poorly drained areas on the High Plains surface at the present time. A maximum depth of leaching of 15 feet has been observed in the Brady soil in Doniphan County, but in northwestern Kansas it is leached of calcium carbonate to depths of only 1½–3 feet. Bignell silt has not been recognized in southwestern Kansas, but the soil developed in the top of Peoria loess is at some places overlain by dune sands (loc. 80). Soil in this stratigraphic position has been referred to as Brady, but as the dune sand may not be an age equivalent to the Bignell loess and as the upper or Tazewellian faunal zone has not been recognized in the Peoria of the area, the soil may have been formed through a longer or shorter time interval than typical Brady farther north in the Plains region. The somewhat thicker profile and more mature development of the so-called "Brady soil" plus the absence of the upper Peoria faunal zone suggest a longer period of soil formation in southwestern than in northern Kansas, even though slight differences in profile morphology might be explained by climatic differences.

LATE WISCONSINAN

The record of late Wisconsinan time in the Kansas region is far less complete and widespread than that of early Wisconsinan time. This may be the result of less vigorous or extensive contributions of outwash by the Caryan-Mankatoan glaciers than from the Iowan-Tazewellian glaciers, to climatic changes, or to other factors unknown. Adequate data are not as yet available concerning the details of advance and retreat of the several late Pleistocene ice sheets of the upper Missouri Basin or the central Rocky Mountains to permit a full appraisal of these factors. Data from molluscan faunas indicate that climate, fauna, and flora during the period of Bignell loess deposition closely resembled present conditions and that the climate was slightly more dry than during the time of Peoria deposition. One aspect of Wisconsinan history is clearly apparent from the stratigraphic sequence in Kansas—namely, that the interval was characterized by two relatively continuous episodes of deposition separated by an interval of comparative stability and soil formation. A similar distribution of glacial events has been suggested for Iowa by Ruhe (1950, p. 441) as the result of the study of drift topographies. Because the Peoria loess is determined to be Iowan and Tazewellian in age, the soil-forming period (Bradyan) is post-Tazewellian. Placement of the overlying Bignell loess in the glacial time scale can be made only by inference. It is not known whether the period of deposition of this latest loess is Caryan, Mankatoan, or both. Comparisons of the morphology of the Brady soil profile with that of the modern profile in similar situations suggest that the time during which the Brady soil formed was as long as the time during which the modern soil formed on the Bignell loess. During late

Wisconsinan time, terraces along the valleys were modified, and the formation of the complex of floodplain terraces was initiated.

BIGNELL MEMBER

The Bignell loess was named by Schultz and Stout (1945, p. 241) from a locality along the south bluff of the Platte River Valley, near the town of North Platte, Nebraska (pl. 1, *A* and *B;* loc. 93). Thin, discontinuous deposits of Bignell are distributed widely over the northwestern quadrant of Kansas. In that area the member is generally less than 10 feet thick, resembles the underlying Peoria in lithology, and contains a distinctive, though sparse, molluscan fauna (fig. 5). Detailed study in Norton County (Frye and A. R. Leonard, 1949, pl. 3) shows that the Bignell is the surficial material in a significant percentage of the upland areas.

Bignell loess has been recognized in the northern Kansas area between Phillips and Doniphan counties only along the bluffs of Republican River Valley. In Doniphan County it occurs discontinuously along the Missouri River bluffs (locs. 1 and 2) and contains a relatively dense concentration of fossil snail shells (fig. 5). In northeastern as well as northwestern Kansas the Bignell seemingly is localized at relatively low places on the Brady soil surface (loc. 1), as some of the highest of the loess hills contain Peoria loess on their summits.

Bignell loess has been recognized at only a few places (loc. 70) in central Kansas, particularly McPherson and Rice counties (Fent, 1950*b*). Elsewhere in the state Bignell loess has not been recognized, but in south-central and southwestern Kansas dune sands at some places occupy the stratigraphic position of the Bignell.

Species	Locality Number							
	1	2	11	19	21	50	60	93
Anguispira alternata	●							
Gastrocopta armifera	●	●						
Hendersonia occulta	●							
Stenotrema monodon aliciae	●	●						
Succinea grosvenori	●							
Succinea ovalis	●	●						
Triodopsis multilineata	●	●						
Vallonia gracilicosta	●							
Vallonia pulchella		●	●		●		●	●
Hawaiia miniscula	●	●	●	●		●		●
Helicodiscus parallelus	●	●		●		●		
Physa anatina						●		
Succinea avara			●		●	●	●	
Total number species	10	7	3	2	2	4	2	2

FIG. 5.—Fossil mollusks from the Bignell loess at seven localities in Kansas and Nebraska

MODERN SOIL

The parent-material of the modern surface soil throughout at least one-third of the area of Kansas is loess of late Pleistocene age. If the areas of the sand-dune tracts of equivalent age are included, at

least half the soils of the state are developed on eolian deposits. Perhaps as much as half of the remaining area contains soils developed on other types of Pleistocene deposits, including alluvium in the many valleys. It is thus apparent that a discussion of the soils that occur in the top of the Pleistocene deposits of Kansas would be a pedological review of much of the state—such an undertaking is clearly beyond the scope of this study. It should be noted, however, that the modern soil developed on Bignell loess is generally shallower and less mature than the underlying Brady buried soil in the same areas and that the depth of profile of each is greatly exceeded by the buried Sangamon (Loveland) soil.

ACKNOWLEDGMENTS.—Work on the Kansas Pleistocene is being carried on by a relatively large group of geologists and is being approached from many aspects. Personnel of the state and federal geological surveys in Kansas who have contributed importantly to the present work include Alvin R. Leonard, Norman Plummer, Ada Swineford, Kenneth L. Walters, Delmar W. Berry, O. S. Fent, and Glenn C. Prescott. James Thorp, W. I. Watkins, William Johnson, and others of the state and federal soil survey have contributed significantly by field conferences and data furnished to us. Personnel of the Iowa Geological Survey, Nebraska Geological Survey, and Nebraska State Museum have given invaluable aid to the work of correlation of the Kansas section with deposits in Iowa and Nebraska by directing us to localities and discussing problems with us in the field. Examination of deposits and collection of fauna from localities in Illinois were made possible by field conferences directed by M. M. Leighton and staff of the Illinois Geological Survey. The manuscript of this paper was improved by the criticisms of R. C. Moore, Ada Swineford, and Alvin R. Leonard.

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PETROGRAPHY OF THE PEORIA LOESS IN KANSAS[1]

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ABSTRACT

The surficial material throughout one-third of Kansas is the Peoria silt (loess) member of the Pleistocene Sanborn formation. Samples of this loess from an area 400 miles east-west by more than 200 miles north-south, which have been analyzed for particle size, chemical composition, and mineral constituents, show that major constituents of the silt-size fraction are quartz, feldspars, volcanic-ash shards, carbonates, and micas; quartz constitutes more than half the volume. A large suite of secondary minerals is present. The clay fraction consists of montmorillonite, illite, calcite, quartz, and feldspar, with a trace of a kaolinite mineral. Evidence used to determine source areas includes (1) progressive linear change in texture; (2) geographic variations in thickness; and (3) minor geographic differences in composition. The Platte River Valley is identified as the primary source for the Kansas region, and the Arikaree, Arkansas, and Republican valleys as important supplementary sources. The Missouri is the predominant source in the northeastern corner of Kansas. Other principal valleys, which lacked glacial outwash, were not source areas of Peoria loess in Kansas.

INTRODUCTION

A deposit of massive homogeneous silt, classed stratigraphically as the Peoria silt member of the Sanborn formation, constitutes the surface or near-surface material in approximately one-third of Kansas. The Peoria member is early Wisconsinan (Pleistocene) in age and consists predominantly of loess. It is the most important and extensive parent-material of surface soils in the state, it constitutes a widely usable ceramic raw material (Frye *et al.*, 1949), and it is extensively used as earth fill in construction. This paper describes the petrography and discusses the origin of the Peoria silt in Kansas.

The Peoria loess of Kansas has its maximum development in the northwestern quarter of the state, where it mantles nearly all upland and intermediate levels to depths ranging from a few feet to 90 feet. It is thin but commonly present in the southwestern and northeastern areas; the maximum observed thickness occurs along the Missouri River bluffs in the extreme northeastern corner of the state. Very thin discontinuous patches occur on some upland areas of south-central and southeastern Kansas, but in these areas the Peoria is included entirely within the profile of the modern topsoil.

In order to attain essential uniformity of petrographic study throughout the state, samples of Peoria silt were taken from Leonard's (1951) intermediate faunal zone (where sections were sufficiently fossiliferous for its identification) or from the approximate middle of the unit. With the possible exception of the localities of very thin loess in Brown, Marshall, and Nemaha counties, all samples are from stratigraphic positions well below a recognizable "B horizon" of the surface soil or of the buried Brady soil, where the Peoria is overlain by the Bignell silt member of the Sanborn formation. Although some of the samples studied have a relatively low calcium carbonate content, they have been essentially unaltered by soil-forming processes since the close of Peoria deposition.

The stratigraphy of the Sanborn formation (Elias, 1931, pp. 163–180) and of the Peoria silt member in the Great Plains region has been described (Lugn, 1935; Schultz and Stout, 1945, 1948;

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Condra, Reed, and Gordon, 1947; Frye and Fent, 1947), and the late Pleistocene loesses, their molluscan fauna, and correlation in the Kansas region are discussed in a companion paper (Frye and Leonard, 1951). The age of the Peoria silt member is early Wisconsinan, and the zone within the member selected for particular petrographic study may be late Iowan or early Tazewellian in age (Leonard, 1951).

METHODS OF STUDY

Field studies of the late Pleistocene deposits in Kansas have been under way for the last ten years as part of several investigations sponsored by the State Geological Survey of Kansas. Chief among these are co-operative ground-water studies conducted by the state and federal surveys, investigations of ceramic materials, and stratigraphic and paleontologic work by the division of basic geology. Several publications resulting from this work contain textural, chemical, and other data on the physical properties of the Peoria loess in Kansas (particularly, Frye, 1945; Swineford and Frye, 1945; Frye and Fent, 1947; Frye *et al.*, 1949; Frye and A. B. Leonard, 1949; Frye and A. R. Leonard, 1949).

SAMPLING

Samples of Peoria loess used in this study were collected during the field seasons of 1948 and 1949 by us and by Norman Plummer and A. Byron Leonard. One- to 25-pound samples were obtained from surface exposures or from auger holes. Surface samples were collected only from localities where stratigraphic relationships were clearly determinable, where exposures were not contaminated by slump or surface weathering products and regional relationships were known. In some localities the stratigraphy was checked by near-by auger holes. Samples for thin sectioning were obtained by shaping a column of loess in place in the exposure and fitting over it a rigid container on which horizontal and vertical orientations were then marked. Auger samples were obtained by the use of a 3-inch spoon-type soil auger which was operated to depths of as much as 24 feet. Although usable data on thickness and stratigraphy of loess, particularly in northern Kansas, were obtained by the portable hydraulic-rotary drilling equipment operated by the Ground-Water Division of the state and federal geological surveys, the samples obtained by hydraulic rotary drilling were judged to be unsuitable for petrographic study and were not used for this purpose.

LABORATORY PROCEDURE

Separate cuts of a single sample were made for each preparation. Laboratory study included chemical and spectrographic analysis (Frye *et al.*, 1949), particle-size analysis, microscopic study of both grain mounts and thin sections, separation and examination of heavy-mineral fractions, electron-microscope examination of clay fractions, and X-ray diffraction analysis of clay fractions.

Particle-size analysis.—Many mechanical analyses of Kansas Peoria loess have been made; but, to assure uniformity of treatment, all data used here were obtained during 1949 by standardized methods. Mechanical analyses were made by the pipette method. The samples were treated with warm dilute (N/10) hydrochloric acid until effervescence stopped. The acid and soluble salts were then removed by filtration through Whatman No. 50 filter paper in a Buchner funnel, and the loess was rinsed several times with distilled water and finally with acetone. After being dried and weighed, the

material was dispersed in distilled water with ammonium hydroxide and agitated for 12 hours in a wrist-action laboratory shaker.

Examination of thin sections indicated that much of the carbonate is secondary or consists of parts of organic structures such as snail shells, although detrital carbonate grains were found to occur in some samples. Furthermore, detrital carbonates presumably have been removed from some samples (particularly from eastern and central Kansas) by weathering processes. It was concluded that more uniform size data, particularly in regard to primary particle size, could be produced by removal of all the carbonate. Microscopic examination of detrital grains in the coarser silt sizes, after completion of mechanical analysis, showed that some of the clay coating of grains had not been removed. Therefore, the finer fractions do not record the total clay content.

Microscopic study.—Oriented thin sections from seven localities (samples 12, 25, 36, 41, 42, and other samples from Thomas County and northern Norton County) were examined (pl. 1). From four of these localities sections were cut both horizontally and vertically, and vertical sections were oriented both east-west and north-south. Heavy minerals were separated from the 62–30 μ size by use of bromoform in a centrifuge, and both the "heavy" and the "light" fractions were examined.

Determination of clay fraction.—Preparations of the finer-than-1 μ fraction of samples from fourteen localities (samples 3, 5, 10, 12, 14, 17, 25, 27, 32, 33, 34, 38, 41, and 42) were examined in the electron microscope. The instrument used is an RCA-EMU-2B. Electron micrographs were made by C. C. McMurtry, department of oncology, University of Kansas School of Medicine (pls. 2 and 3). The clay particles were dried from a drop of dilute water suspension on a collodion film supported by a 200-mesh stainless-steel specimen screen. They were then shadow-cast with 175 Å of chromium at an angle of 1:5 from the plane of the film (producing shadows five times the height of the particles) and 15 Å of aluminum was deposited vertically. The magnification in the microscope was 7,800×.

X-ray diffraction spectrometer patterns of the less-than-1 μ fraction of samples 12 and 42 were obtained through the courtesy of Mr. J. W. Ranftl, G.E. X-Ray Corporation, Milwaukee. Camera patterns of samples 14 and 38 were made and interpreted by Dr. R. E. Grim, Illinois State Geological Survey.

PETROGRAPHY

TEXTURE

Particle-size data for 47 samples from 42 localities are shown in table 1. The size-frequency distributions are similar to those obtained for samples classed as loess from elsewhere in North America (Smith, 1942; Kay and Graham, 1943;

PLATE 1

Photomicrographs of thin sections of Kansas Peoria loess. Plane-polarized light.

A, Sample 12, Rawlins County, Kansas; 75×. Note coarse silt particles and large volcanic ash shards.

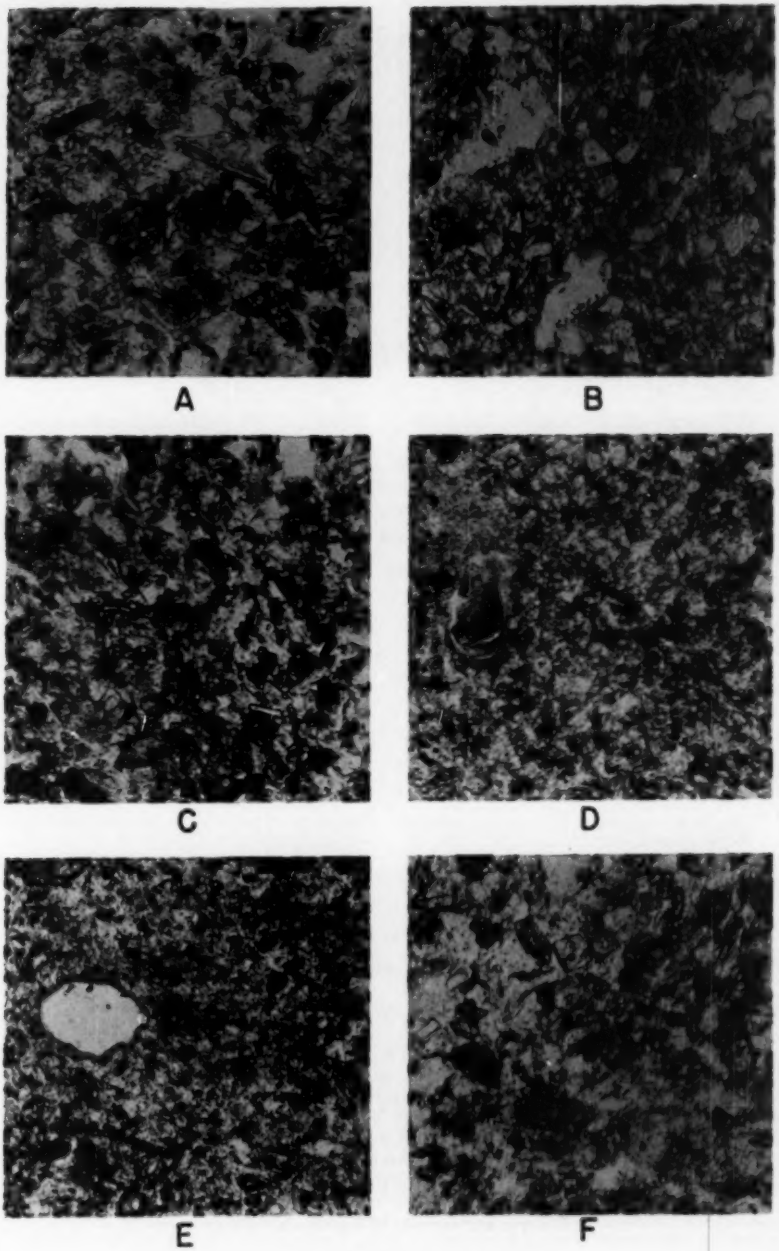
B, NW. $\frac{1}{4}$ of NW. $\frac{1}{4}$, Sec. 26, T. 2 S., R. 23 W., Norton County, Kansas; 75×. Volcanic ash shards common.

C, Sample 25, Norton County, Kansas; 75×. Particle size similar to that in *B*.

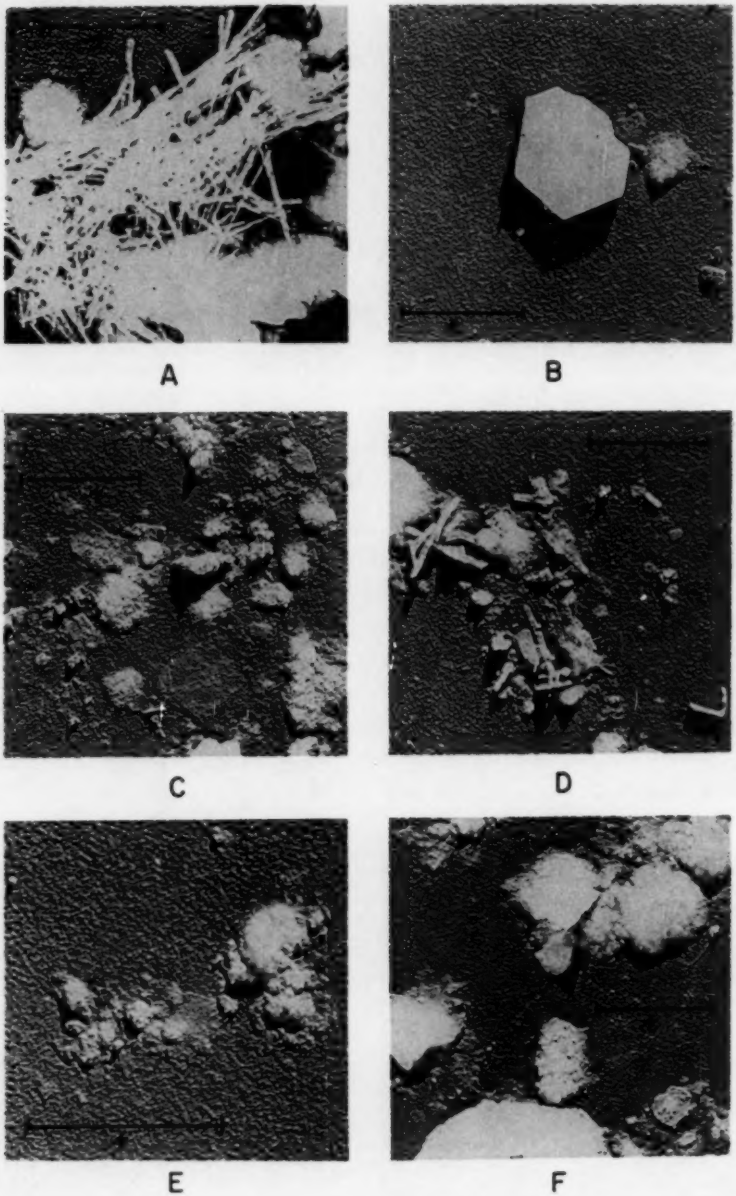
D, Sample 36, Republic County, Kansas; 75×. Note clay filling hole.

E, Sample 41, Brown County, Kansas; 75×. Note very fine particle size and clay-lined hole.

F, Sample 42, Doniphan County, Kansas; 150×. Silt-sized dolomite grains common.



Photomicrographs of Kansas Peoria loess



Electron micrographs of Kansas Peoria loess

Russell, 1944) and as wind-blown silt (Swineford and Frye, 1945). The median diameter ranges from 4.29ϕ (0.051 mm.) in sample 17 to 5.59ϕ (0.021 mm.) in Brown County. The sorting, as shown by the phi quartile deviation, ranges from 0.36ϕ in Lincoln County, Nebraska, to 0.89ϕ in Nemaha and Stanton counties, if the exceptional Finney County locality ($QD_\phi = 1.13$) is ignored. The Finney County material is judged to be a mixture of a traction load of sand and a suspension load of silt.

The range in median diameter is only slightly more than one Udden grade, but it displays a high degree of geographic consistency. The coarsest samples were those from localities adjacent to the Arkansas, Platte, Republican, and Missouri River valleys. Textural traverses are presented in figure 1. Traverses *A* and *B* show progressive decrease to the south and southeast away from the Arikaree River valley, and traverses *D* and *F* show a comparable progressive decrease in median grain size southward from the Platte. Peoria loess is coarse-textured in the northwestern part of the northwestern quadrant of Kansas and becomes finer along all traverses toward the south and east (fig. 2). A more rapid decrease in grain size is shown eastward from the east bluff of the Republican Valley in Republic County and westward from the west bluff of Missouri Valley in Doniphan County (fig. 1, *C*). Loess of finest texture was found adjacent to the west side of Republican Valley in Republic County and midway between the Republican and Missouri rivers in Nemaha and Brown counties. In general, the Peoria loess has the coarsest texture where the member is thickest (fig. 3).

A definite relationship is seen between particle size and degree of sorting of the loess samples, the coarser samples being generally better sorted than the finer ones. Griffiths (1948) showed that in some Trinidad sediments association between the phi median diameter (Md_ϕ) and the sorting (his phi percentile deviation, $PD_\phi = [P_{10\phi} - P_{90\phi}]/2$) measured by the correlation coefficient gives values of r from 0.59 to 0.79. A value of $r = 0.49 (P < 0.01)$ is obtained for the present loess data. If the sandy Finney County material (sample 17) is omitted, r becomes 0.65, which is well within the range of r for Griffiths' Trinidad sediments.

PLATE 2

Electron micrographs of the less-than-1-micron fraction of Kansas Peoria loess. Electron micrographs by C. C. McMurtry.

A, Sample 10, Stanton County, Kansas; 19,000×. Rods judged to be calcite intermixed with montmorillonite-type particles.

B, Sample 25, Norton County, Kansas; 19,000×. Kaolinite plate.

C, Sample 33, Republic County, Kansas; 19,000×. Illite-like and montmorillonite-like particles. This sample is from west of Republican River, where the Peoria is thin and weathered; note contrast with *D*.

D, Sample 34, Republic County, Kansas; 19,000×. Sample from east bluff of Republican Valley, where Peoria loess is 40 feet thick; note abundant rods judged to be calcite. The elongate particles are shown to be rods rather than laths by the length of the chromium shadows.

E, Sample 27, Smith County, Kansas; 31,000×. Typical of Peoria loess where it is thin in the central and eastern areas.

F, Sample 42, Doniphan County, Kansas, from bluff of Missouri River Valley; 19,000×. Montmorillonite and illite typical of the northeastern area; note absence of rod-shaped particles. X-ray diffraction data for this size fraction indicate abundant montmorillonite and illite but no calcite.

STRUCTURE

The Peoria loess has a massive appearance in outcrop sections except at a few localities immediately adjacent to the Platte, Republican, and Missouri valleys, where distinct lamination is observed. In the northeastern Kansas area, but not in the western part of the state, brown- to buff-colored bands or laminae, in some places displaying a contorted pattern, are common, as are elongate vertical limonite concretions suggestive of accumulation around rootlets. All thin sections studied were distinguished by lack of laminae, the presence of small elongate and circular cavities, no preferred orientation of the micas, and in those from northeastern Kansas by limonite mottling. Many of the cavities are lined by yellowish-brown, highly birefringent clay. In the western part of the state small calcium carbonate concretions, 1 mm. or less in diameter, are common. These are well below the zone of caliche accumulation of the soil which typically displays much larger concretions.

Gross structural features show a marked contrast between western and eastern Kansas loess deposits. Along the Missouri Valley bluffs (Doniphan County), the Peoria displays persistent vertical cleavage on vertical faces, and vertical faces of road cuts tend to retain their verticality for many years; the entire unit displays a massive cohesion, and slumpage is commonly characterized by large blocks. Strongly developed vertical cleavage is rarely observed in northwestern Kansas, persistence of verticality of cut faces is less common, and slopes on underlying formations are generally veneered by silt that has moved downward by creep processes (rather than as slump blocks) from the higher loess units.

COMPOSITION

Silt fraction.—Within the silt-size fraction, the major constituents were determined, from thin sections and grain mounts, to be quartz, feldspars, volcanic-ash shards, carbonates, and micas. Quartz makes up slightly more than half the total volume of the deposit. The mineral inclusions and bubble chains suggest that most of the quartz is of igneous origin, although as much as 10–20 per cent may be metamorphic. The grains are subangular to angular and are free from secondary crystal growths.

Feldspars, which are the second most abundant mineral group, occur generally in a ratio to quartz of 1:4. Orthoclase is the prevalent feldspar, and sodic plagioclase (*ca.* $Ab_{80}An_{20}$) and microcline are generally present. More calcic plagio-

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PLATE 3

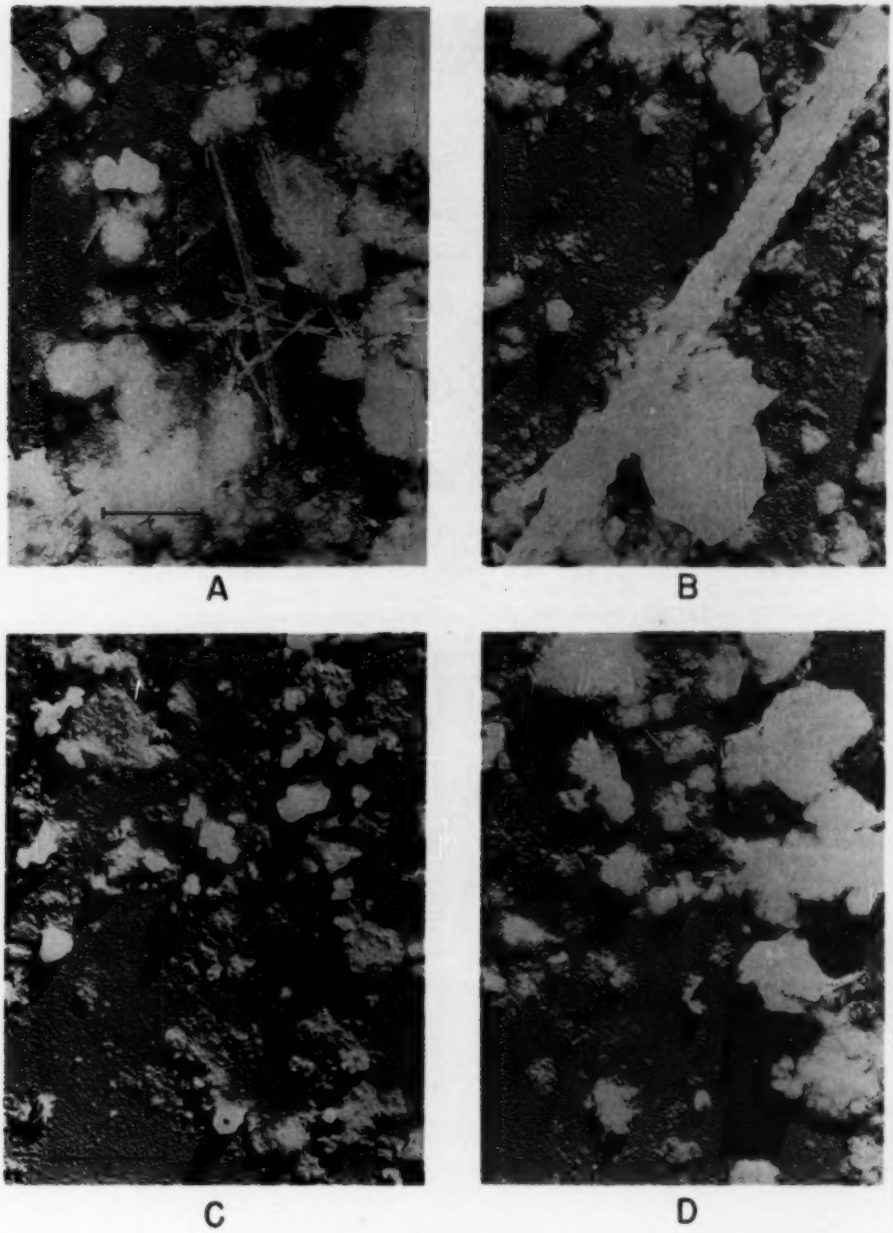
Electron micrographs of the less-than-1-micron fraction of Kansas Peoria loess. Electron micrographs by C. C. McMurtry.

A, Sample 12, Rawlins County, Kansas; 10,000×. Cottony aggregates of montmorillonite, thin flakes of illite, and elongate lathlike structures of unknown origin, presumably calcite. The presence of montmorillonite, illite, and calcite in the sample was confirmed by X-ray diffraction data.

B, Sample 25, Norton County, Kansas; 11,000×. Tube structure of unknown origin found only in sample from this locality.

C, Sample 38, Marshall County, Kansas; 11,000×. Montmorillonite, illite, and dense equidimensional particles having octahedral habit, judged to be cristobalite. Note well-developed octahedron and shadow near right end of micron mark.

D, Sample 32, Rice County, Kansas; 11,000×. Montmorillonite, illite, and a few particles that may be quartz or feldspar.



Electron micrographs of Kansas Peoria loess

TABLE 1

MECHANICAL ANALYSES OF MIDDLE PEORIA LOESS

No.	Location (Sec., T., R.)	County	Size Distribution (Per Cent by Weight)											M_ϕ	σ_ϕ	Md_ϕ	OD_ϕ	Sk_ϕ	PD_ϕ	Per Cent Loss (N/10 HCl)
			Phi Units	1–2	2–3	3–4	4–4.5	4.5–5	5–6	6–7	7–8	8–9	9–10	10+						
			Microns	500–250	250–125	125–62	62–44	44–31.2	31.2–15.6	15.6–7.8	7.8–3.9	3.9–1.95	1.95–1	<1						
	Nebraska																			
1a	9, 7 N., 29 W.	Frontier			2.8	20.2	32.4	21.5	15.8	3.8	1.8	0.4	0.4	1.0	4.63	1.138	4.42	0.46	0.554	1.16	6.0
1b	9, 7 N., 29 W.	Frontier			1.8	18.9	32.2	22.6	16.4	4.2	1.5	0.6	0.4	1.3	4.69	1.164	4.45	0.46	0.619	1.16	8.7
1c	9, 7 N., 29 W.	Frontier			0.4	15.4	31.2	25.9	18.8	4.0	1.4	0.4	0.6	1.8	4.80	1.189	4.56	0.44	0.606	0.94	9.2
2	E. ½ E. ½ 3, 12 N., 29 W.	Lincoln			0.8	14.9	38.2	26.1	14.3	2.8	1.6	0.2	0.4	0.8	4.66	0.987	4.45	0.36	0.638	0.93	8.9
	Kansas																			
3a	SW. SE. 17, 1 S., 41 W.*	Cheyenne			1.1	15.2	36.0	21.1	16.0	4.3	2.1	1.6	0.2	2.5	4.86	1.339	4.47	0.48	0.874	1.15	13.4
3b	SW. SE. 17, 1 S., 41 W.*	Cheyenne			0.6	12.7	42.9	20.2	10.7	4.1	2.7	1.7	1.2	3.2	4.95	1.478	4.43	0.40	1.055	1.44	10.9
3c	SW. SE. 17, 1 S., 41 W.*	Cheyenne			2.0	11.7	29.8	20.6	25.3	5.5	1.8	0.8	1.0	1.4	4.93	1.238	4.66	0.56	0.654	1.14	13.2
4	SE. SE. 28, 3 S., 39 W.	Cheyenne				4.2	36.8	29.7	18.4	4.5	2.1	0.5	0.9	2.8	5.03	1.276	4.65	0.41	0.893	1.01	13.1
5	SW. SW. 31, 9 S., 39 W.	Sherman				0.7	29.9	28.1	27.2	6.6	3.2	0.9	1.1	2.3	5.22	1.248	4.85	0.54	0.889	1.22	15.1
6	SW. SW. 23, 12 S., 40 W.	Wallace			1.1	2.8	29.6	26.1	25.9	6.8	3.1	1.0	1.2	2.3	5.18	1.309	4.82	0.56	0.825	1.17	14.3
7	SW. NW. 9, 16 S, 40 W.	Greeley			1.0	5.1	23.6	21.6	27.6	10.5	4.6	2.2	1.8	2.0	5.35	1.414	4.97	0.69	0.806	1.50	17.3
8	SW. NW. 19, 21 S., 40 W.	Hamilton			0.5	7.4	16.8	22.0	29.3	12.1	5.4	2.8	1.2	2.4	5.43	1.451	5.11	0.72	0.662	1.62	13.4
9	NE. SE. 12, 25 S., 41 W.	Hamilton			2.4	12.1	19.5	22.9	27.4	8.3	3.7	1.2	0.4	1.9	5.06	1.347	4.85	0.64	0.468	1.40	15.8
10	SE. 25, 29 S., 43 W.	Stanton		1.4	3.8	12.1	12.3	16.1	27.3	13.0	5.9	3.3	2.2	2.6	5.36	1.737	5.19	0.89	0.294	2.10	17.7
11	CEL. NE. 29, 3 S., 36 W.	Rawlins			1.1	33.7	33.1	20.6	6.7	1.3	0.9	0.2	2.4	5.06	1.161	4.73	0.44	0.853	1.00	12.2
12	SW. SW. 2, 3 S., 33 W.	Rawlins			0.2	9.5	30.5	28.2	22.5	4.3	2.5	0.2	0.6	1.4	4.91	1.115	4.67	0.48	0.646	0.96	12.3
13	SW. 31, 6 S, 33 W.	Thomas			1.6	3.1	28.9	28.0	23.2	7.2	2.8	1.7	1.7	1.7	5.15	1.323	4.79	0.56	0.816	1.28	12.0
14	NW. NE 4, 11 S., 30 W.	Gove			tr	4.6	26.5	24.5	27.6	8.3	3.7	1.8	0.9	2.2	5.26	1.312	4.89	0.62	0.846	1.30	13.6
15	SE. 6, 17 S., 32 W.	Scott				0.9	22.5	26.3	27.7	10.3	5.4	2.6	1.6	2.6	5.48	1.408	5.01	0.68	1.001	1.58	20.3
16	SE. SE. 35, 20 S., 35 W.	Wichita				0.9	24.0	21.8	30.6	11.5	5.3	2.4	1.5	2.0	5.48	1.333	5.11	0.69	0.833	1.49	15.2
17	SE. 4, 24 S., 32 W.	Finney	2.2		21.4	19.5	11.8	13.0	17.6	7.6	3.2	1.2	0.8	1.8	4.43	1.747	4.29	1.13	0.240	2.04	13.8
18	NW. NW. 6, 28 S., 32 W.	Haskell	0.3		1.2	5.5	16.1	18.9	33.1	14.4	5.4	2.3	1.2	1.6	5.44	1.372	5.24	0.72	0.437	1.46	16.9
19	CWL. SW. 15, 6 S., 28 W.	Sheridan				4.5	28.7	25.9	25.7	7.3	2.8	2.1	0.6	2.4	5.20	1.306	4.82	0.58	0.873	1.27	9.8
20	NE. 9, 8 S., 28 W.	Sheridan				1.3	26.1	23.0	28.7	10.0	3.9	2.4	2.2	2.4	5.44	1.401	4.99	0.67	0.964	1.49	11.9
21	NE. NE. 12, 3 S., 27 W.	Decatur				1.8	30.3	26.8	26.6	6.3	2.8	2.0	0.8	2.8	5.25	1.326	4.83	0.57	0.950	1.24	8.6
22	SW. SE. 5, 12 S., 24 W.	Trego			tr	0.6	18.0	24.5	33.4	12.6	5.5	2.3	1.3	1.9	5.52	1.298	5.21	0.64	0.716	1.40	11.9
23	NW. NW. 26, 23 S., 25 W.	Hodgeman	tr		0.2	1.9	13.0	19.5	37.4	15.7	6.5	2.5	0.8	2.5	5.65	1.338	5.41	0.69	0.528	1.52	19.4
24	NE. NW. 6, 30 S., 23 W.	Clark	0.8		1.5	12.4	20.5	19.9	26.4	11.3	3.4	1.6	0.7	1.6	5.11	1.384	4.87	0.74	0.520	1.46	21.3
25	SW. NW. 11, 3 S., 23 W.	Norton				0.1	25.7	28.5	30.0	9.3	2.2	1.5	0.6	1.9	5.26	1.173	4.93	0.56	0.844	1.13	11.4
26	NE. SE. 4, 2 S., 19 W.	Phillips				0.8	20.8	24.9	26.8	13.2	4.8	2.4	2.2	4.3	5.63	1.561	5.13	0.78	0.961	1.74	9.2
27	NW. NW. 18, 2 S., 13 W.	Smith				1.3	22.5	27.6	31.4	8.9	3.6	1.9	0.4	2.3	5.32	1.251	4.97	0.58	0.839	1.24	10.6
28	SE. NW. 4, 1 S., 7 W.	Jewell				3.9	23.2	28.9	31.5	7.7	2.0	0.5	0.3	2.0	5.14	1.141	4.90	0.52	0.631	1.04	7.4
29	NE. SW. 25, 3 S., 8 W.	Jewell				1.1	19.4	24.3	35.5	11.4	4.0	2.1	0.2	2.1	5.41	1.222	5.15	0.60	0.638	1.29	5.9
30	SE. NE. 29, 7 S., 7 W.	Mitchell	0.3		tr	tr	19.1	27.6	32.5	12.5	3.8	1.1	0.7	2.4	5.40	1.262	5.09	0.61	0.737	1.25	9.8
31	NW. 6, 11 S., 7 W.	Lincoln				tr	16.3	20.4	36.9	11.4	4.7	2.0	2.0	6.1	5.78	1.617	5.36	0.68	0.779	1.84	7.0
32	NW. SW. 7, 18 S., 7 W.	Rice	0.2		0.6	14.3	29.2	21.6	19.2	7.6	3.4	1.2	1.2	1.6	4.99	1.348	4.63	0.60	0.801	1.42	9.6
33	NE. NE. 15, 3 S., 5 W.	Republic				0.4	11.9	21.4	38.3	14.8	5.8	3.1	1.6	2.7	5.72	1.376	5.43	0.68	0.632	1.52	6.4
34a	NE. NW. 5, 3 S., 4 W.†	Republic			0.1	7.5	31.8	30.7	19.7	4.3	2.0	0.6	1.4	1.8	4.97	1.226	4.67	0.43	0.734	1.00	13.8
34b	NE. NW. 5, 3 S., 4 W.†	Republic				4.2	30.2	37.2	22.3	2.7	1.5	0.5	0.6	0.8	4.88	0.940	4.71	0.37	0.543	0.82	7.7
35	SW. SW. 35, 2 S., 3 W.	Republic				0.6	10.7	23.4	38.9	15.7	5.6	2.1	1.0	2.1	5.62	1.266	5.39	0.63	0.545	1.35	12.6
36	NW. NW. 9, 3 S., 1 W.	Republic				0.7	14.0	22.6	35.4	15.4	4.1	2.4	1.7	3.7	5.68	1.446	5.36	0.68	0.664	1.49	10.6
37	NW. 21, 1 S., 5 E.	Washington				0.2	11.4	20.6	40.6	15.7	5.5	2.5	1.4	2.0	5.67	1.275	5.44	0.62	0.541	1.37	8.5
38	SE. SW. 13, 1 S., 7 E.	Marshall				tr	10.8	20.3	39.8	17.0	6.0	2.7	1.0	2.3	5.71	1.286	5.47	0.65	0.500	1.40	10.4
39	NW. 19, 1 S., 10 E.	Marshall				0.2	20.0	16.2	37.7	14.4	5.9	1.8	2.4	1.4	5.62	1.278	5.36	0.68	0.610	1.50	10.4
40	NE. NW. 33, 2 S., 11 E.	Nemaha		0.1		0.2	13.7	17.8	32.4	15.4	6.0	3.4	3.1	7.9	6.07	1.782	5.56	0.89	0.859	2.47	7.4
41	SE. SE. 23, 2 S., 17 E.	Brown				tr	7.5	17.3	42.6	17.7	4.8	2.5	0.4	7.3	5.98	1.580	5.59	0.66	0.741	1.74	8.1
42	NE. SE. 6, 2 S., 20 E.	Doniphan				0.6	17.6	30.2	34.1	5.6	2.1	0.5	1.2	2.1	5.22	1.176	4.94	0.44	0.663	0.92	10.2

* *a*, 80 feet above base; *b*, 70 feet above base; *c*, 50 feet above base.

† *a*, 10 feet and 11 inches above base; *b*, 30 feet above base.

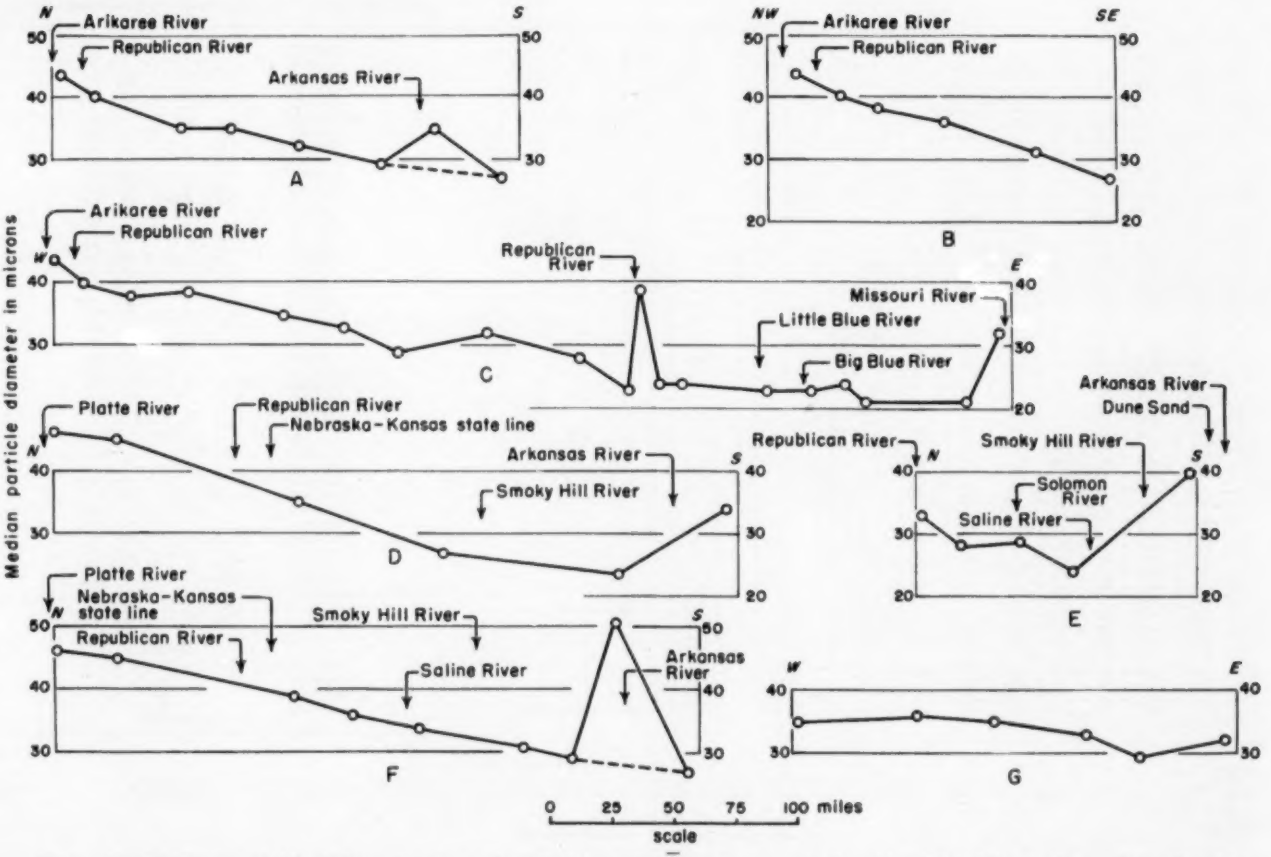


FIG. 1.—Textural traverses of Peoria loess. Samples are from the intermediate faunal zone or stratigraphic middle of the Peoria member; values determined by pipette analysis are reported as median diameter in microns. Note relatively uniform decrease in median diameter away from principal sources, and secondary maximums at subsidiary sources.

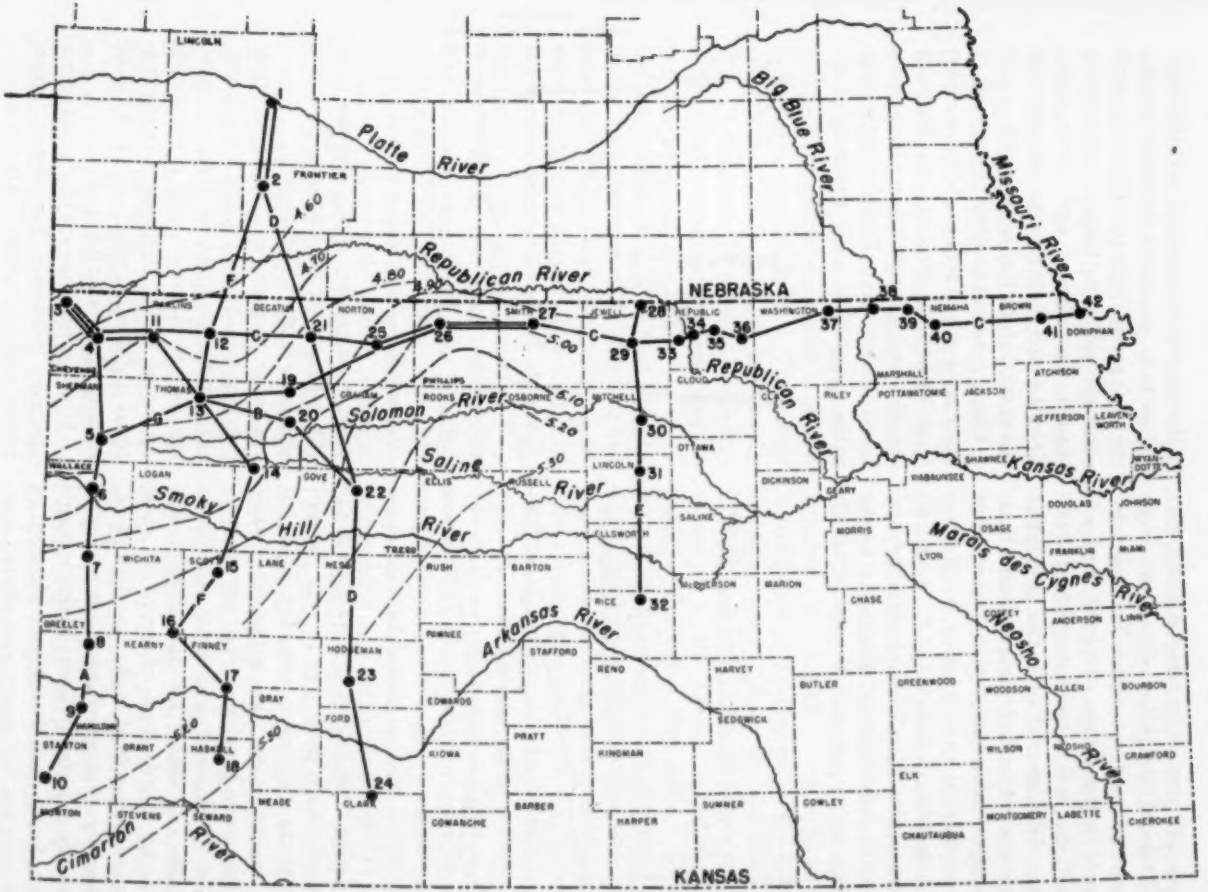


FIG. 2.—Map of Kansas and southern Nebraska showing localities where Peoria loess was sampled and sample numbers used in this report. The position and number of textural traverses are shown. Dashed lines in northwestern Kansas show generalized equivalence of phi median diameter.

clases are present in some samples, but are rare. Clear grains of sanidine were observed in a few samples. Feldspars decrease in percentage eastward, and microcline is rare in the eastern third of the state. All degrees of alteration of feldspar grains occur in each sample studied, although the eastern samples show a higher percentage of altered grains.

Volcanic-ash shards (pl. 1, *A* and *B*) were observed in all samples studied, but their abundance ranges from a trace to as much as 10 per cent of the coarse silt fraction. The indices of refraction of the shards range from 1.49 to 1.52, in contrast to the more constant index of 1.499–1.502 for the widespread early Pleistocene Pearlette ash of the same region (Swineford and Frye, 1946). In shape the shards are not uniform, but fibrous shards were observed in most samples examined, and a few showed microspherulites which have not been previously reported in Great Plains volcanic ash. Some fresh grains were observed in all samples, but many were altered, particularly at the margins.

Micas generally constitute about 3 per cent of the coarse silt fraction. Muscovite is the most common mica, and the ratio of biotite to muscovite decreases toward the east. Several varieties of chlorite were observed, although they constitute less than 1 per cent of the silt fraction. Chlorites are slightly more abundant in the northeastern Kansas samples.

Chert constitutes about 1 per cent of the coarse silt fraction of the samples from Doniphan, Brown, and Nemaha counties but is rare to absent in samples taken farther west. Unidentified aggregates, possibly of very fine-grained altered metamorphic rocks, are also present in the coarse silt fraction. These are, as a rule, fairly well rounded.

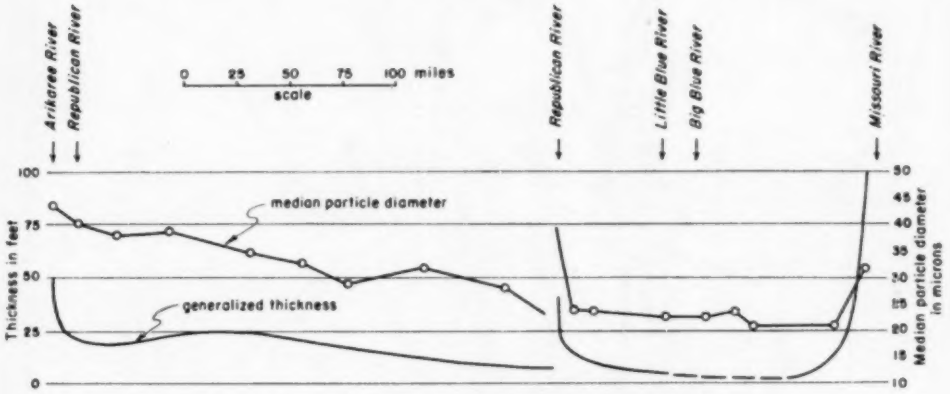


FIG. 3.—Comparison of texture and generalized thickness along traverse *C*

Calcite occurs as large, irregular patches, minute grains, and silt-size particles in the thin sections from the western half of the state. The predominant carbonate in the Doniphan County material is dolomite, which occurs as small rhombs of silt size (pl. 1, *F*).

Minor accessories are black and brown opaques, leucoxene, hornblende (green, brown, and blue-green), chlorite, biotite, epidote, garnet (colorless and pink), various pyroxenes, tremolite-actinolite, zircon, tourmaline, rutile (red and yellow),

staurolite (red-brown prismatic and some yellow), titanite, sillimanite, and zoisite.

Most of the silt grains are coated with a pale-green, highly birefringent clay mineral. After treatment for 24 hours in hot dilute hydrochloric acid the birefringence is destroyed, but the green color remains. A clay coating on grains in the loess from western Kansas, Nebraska, and Colorado was identified by Mielenz, Holland, and King (1949) as montmorillonite.

The percentage loss after digestion in warm dilute hydrochloric acid is shown in the last column of table 1. The values range from 7 to 21 per cent. In general, there is more than 10 per cent loss in the western half of the area and less than 10 per cent in the eastern half.

Clay fraction.—Study of thin sections shows that clay materials coat the silt grains, occur as alteration products in some of the feldspars, coat the walls of the various holes (pl. 1, *E*), fill some cavities (pl. 1, *D*), and serve as matrix.

The fractions finer than 1 μ from 14 samples were examined in the electron microscope. Particles characteristic of montmorillonite and illite were present in all the preparations studied. Samples from the thick unleached loess deposits of the western third of Kansas (samples 3, 6, 10, 12, 14, 17, and 25) were characterized by the presence of varying amounts of elongate fibers—or rod-shaped particles—of unknown origin. Such particles are also abundant in the thick loess (sample 34) immediately east of the Republican River in Republic County, rare in the Doniphan County loess (sample 42), and extremely rare to absent in the other samples examined (pls. 2, *C* and *E*, and 3, *C* and *D*).

The elongate structures are uniform in width and thickness in all samples except the Norton County loess (sample 25) but show great variation in length. The average width is approximately 400 Å. The thickness, as shown by the chromium shadows, is approximately equal to the width (see especially pl. 2, *D*, rod in lower right-hand corner). The length ranges from less than 0.1 to more than 3 μ. The surfaces are smooth, but a few slight irregularities give minor variations in thickness. The particles are not entirely opaque to the electron radiation, except where a considerable thickness of chromium has been deposited on them. The structures occur singly, in small bundles of two to five parallel units (pl. 3, *A*), and as large "nests" (pl. 2, *A*). The Norton County loess also has much larger structures (pl. 3, *B*). These are rod-shaped particles, having a diameter of about 0.5 μ and an average length of about 8 μ. The edges have a ragged appearance.

Other constituents indicated by electron micrographs are a few thin hexagonal flakes, presumably kaolinite (pl. 2, *B*); occasional shards of volcanic ash in a few of the western Kansas samples; a few minute, flat, rhomb-shaped particles in the Wallace County loess (sample 6), suggestive of dolomite or possibly calcite; and small dense octahedral particles in several samples (particularly Marshall County, sample 38) thought to be possibly α-cristobalite. Several octahedral particles are shown in plate 3, *C*.

X-ray diffraction data were obtained for four samples. Spectrometer patterns were made on the less-than-1 μ fraction of samples 12 and 42 by Mr. J. W. Ranftl of the G.E. X-Ray Division in Milwaukee. The pattern for the Rawlins County material (sample 12) indicates the presence of calcite, a montmorillonite mineral, illite, some quartz, and a small quantity of α-cristobalite (pl. 3, *A*). The pattern for the Doniphan County loess suggests the predominance of montmoril-

lonite and illite (pl. 2, *F*), but there are no calcite reflections. Although thin-section examination of this sample shows abundant dolmite in the coarser silt sizes (pl. 1, *F*), there are no dolomite reflections in the diffraction pattern made from the finer material.

Diffraction patterns for the less-than-1 μ fraction of samples 14 and 38 were made and interpreted by Dr. R. E. Grim. He reports (personal communication) that sample 14 (Gove County) contains about 30 per cent montmorillonite, 20 per cent illite, 30 per cent calcite, 15 per cent quartz, some feldspar, and a faint suggestion of a trace of a kaolinite-type mineral. Sample 38 (Marshall County) contains about 60 per cent montmorillonite, 20 per cent illite, 10 per cent quartz, 10 per cent feldspar, and a faint suggestion of a trace of a kaolinite-type mineral. Although no indication of cristobalite was found in the diffraction patterns for sample 14 or sample 38, the sensitivity to this mineral would be about 2–4 per cent, and its presence is indicated by electron-microscope study (pl. 3, *C*).

The absence of prominent halloysite reflections in sample 14, which was estimated by electron-microscope study to contain about 40 per cent of the peculiar elongate particles, rules out the possibility that this material is halloysite. Inasmuch as (1) the particles are totally destroyed by acetic acid; (2) there are no large numbers of other particles having morphology suggestive of calcite visible in the electron microscope; (3) the only abundant crystalline components are shown by X-ray diffraction to be calcite, montmorillonite, illite, and quartz; and (4) their occurrence is restricted to samples from which all the calcium carbonate has not been leached, it is concluded that the small rods are composed of calcite. It is further noted that the less-than-1 μ fraction of sample 14 contains a larger proportion of calcite (30 per cent) than does the sample as a whole, even if it be assumed that the total hydrochloric acid loss (13.6 per cent, table 1) is calcium carbonate. Grim (personal communication) suggests that these rod-like structures may be of organic nature. If so, they may have been produced by fossil organisms, or they may even be of Recent age.

Chemical composition.—Chemical analyses of loess from seven localities were made as part of a study of ceramic utilization (Frye *et al.*, 1949); partial analyses of the Peoria for these localities are given in table 2. In general, these analyses substantiate the petrographic data, indicating a relatively high degree of uniformity throughout the area. The average composition is similar to that of a granite except that the alkalies are relatively low, and ten analyses of TiO_2 (included with Al_2O_3 in table 2) show higher values (1.02 per cent) than are common in granites. Silica is higher in the northeastern Kansas samples than in those from western Kansas, whereas Al_2O_3 and CaO are relatively higher in the west.

ORIGIN

The origin of massive silt deposits (loess) has been a subject of controversy for more than fifty years. Disagreement among workers has generally centered on the agent of transportation and deposition, and, at one time or another, streams, lakes, winds, and processes of mass movement on slopes have all been championed. Extensive deposits of massive silts have been known since the 1890's to blanket the uplands in large areas of northern and western Kansas (Haworth, 1897), and during the past decade these deposits have been studied over much of Kansas by members of the

State Geological Survey staff (Elias, 1931; Frye, 1945, 1949; Swineford and Frye, 1945; Frye and Fent, 1947; Frye, Swineford, and Leonard, 1948; Frye *et al.*, 1949; Frye and A. B. Leonard, 1949). Although some relatively structureless silts on slopes are derived colluvially from higher silt deposits (Elias, 1931) and some loess-like silts at low levels are water-laid, these studies have led to the conclusion that the extensive deposits of massive silts over thousands of square miles of upland and high-terrace surfaces are predominantly the result of eolian action. The facts contributing most importantly to this conclusion are (1) topographic position of loess on extensive divide areas, including highest elements in local topography; (2) textural similarity to that of modern wind-transported silt; (3) relatively uniform composition over an area of 30,000–40,000 square miles in Kansas alone, where the loess rests on various stratigraphic units of Pennsylvanian, Permian, Cretaceous, Pliocene, and Pleistocene age; (4) decrease in thickness and median grain diameter in directions away from major outwash-carrying valleys; (5) regional persistence of distinct faunal zones characterized by terrestrial snails; (6) lateral persistence of buried soil profiles which lack evidence of erosion or creep; (7) gradational contact of overlying loess units on buried soils; and (8) lack of any other known agent capable of depositing uniform silts simultaneously on sharply discordant topographic levels. In view of these facts, an eolian origin of the Kansas Peoria loess is accepted without reservation and is implicit in further discussions of local source and distribution patterns.

TABLE 2

CHEMICAL ANALYSES OF PEORIA LOESS IN KANSAS*

Location	Depth below Top (Feet)	SiO_2	Al_2O_3 and TiO_2	Fe_2O_3	CaO	MgO	K_2O	Na_2O†	Ign. Loss	H_2O ($104°$–$104°$)
NE. ¼ Sec. 32, T. 12 S., R. 37 W.	6	65.37	12.83	3.41	5.88	1.89	2.78	1.14	6.60	0.73
NE. ¼ Sec. 32, T. 12 S., R. 37 W.	10	65.37	13.30	3.10	5.85	1.79	2.89	1.45	6.15	0.86
NE. ¼ Sec. 32, T. 12 S., R. 37 W.	15	67.19	14.45	3.55	3.76	1.99	2.94	1.37	4.75	0.97
SE. ¼ SE. ¼ Sec. 28, T. 3 S., R. 39 W.	4	67.62	12.25	2.86	5.57	1.55	3.47	0.71	5.91	0.42
SE. ¼ SE. ¼ Sec. 28, T. 3 S., R. 39 W.	12	69.99	12.67	2.84	4.21	1.62	2.86	1.15	4.57	0.19
NW. ¼ NW. ¼ Sec. 26, T. 2 S., R. 23 W.	7	68.39	12.92	3.56	3.32	1.79	2.57	1.75	5.54	0.67
NW. ¼ NW. ¼ Sec. 26, T. 2 S., R. 23 W.	11	70.16	14.02	2.70	2.87	2.22	2.19	0.97	4.12	0.67
NW. ¼ NW. ¼ Sec. 26, T. 2 S., R. 23 W.	15	72.35	13.56	3.46	2.04	1.77	2.78	0.24	3.66	0.83
NW. ¼ SW. ¼ Sec. 7, T. 18 S., R. 7 W.	4	72.87	13.25	3.20	0.84	1.11	nd	4.24	4.49	0.40
NW. ¼ SW. ¼ Sec. 7, T. 18 S., R. 7 W.	6	76.66	13.13	3.36	0.91	1.01	nd	1.36	3.47	0.28
SW. ¼ SE. ¼ Sec. 35, T. 2 S., R. 3 W.	3.5	71.01	13.24	3.89	1.52	1.81	2.31	2.28	3.84	0.20
NE. ¼ NE. ¼ Sec. 19, T. 8 S., R. 3 E.	8	75.86	11.62	2.94	1.32	0.89	3.18	1.77	2.31	0.23
NE. ¼ SE. ¼ Sec. 6, T. 2 S., R. 20 E.	10	77.38	11.82	2.72	0.95	1.26	2.69	0.77	2.41	0.40
NE. ¼ SE. ¼ Sec. 6, T. 2 S., R. 20 E.	15	76.79	10.85	2.87	1.10	1.15	2.58	2.39	2.27	0.36
NE. ¼ SE. ¼ Sec. 6, T. 2 S., R. 20 E.	20	73.33	12.00	3.02	2.67	1.06	2.55	1.48	3.89	0.47
NE. ¼ SE. ¼ Sec. 6, T. 2 S., R. 20 E.	25	71.62	11.61	2.92	3.55	2.08	2.00	1.54	4.68	0.56
NE. ¼ SE. ¼ Sec. 6, T. 2 S., R. 20 E.	30	71.21	10.11	2.61	3.70	2.62	2.35	1.62	5.78	0.13
Average		71.36	12.57	3.12	2.94	1.62	2.68	1.54	4.38	0.49

* Analyses by Russell Runnels, State Geological Survey of Kansas. † Determined by difference.

SOURCE

Considering that air currents transported and deposited the extensive

blankets of upland loess in Kansas, there is the problem of determining the areas of principal silt supply within this broad region. As Kansas is far from the margins of late Pleistocene glaciers and as the Peoria of the uplands rests generally on the uneroded surface of a buried soil, the valley flats of the major drainageways present the only available source for eolian silts. The total volume of silt included in the Peoria is large—57 cubic miles in Kansas if we assume an average thickness of 10 feet over 30,000 square miles. This volume of Peoria silt is roughly comparable to the volume of alluvium that would be contained in 600–800 miles of valley similar to the Missouri of northeastern Kansas. This volume of sediment demonstrates that the source of Peoria loess must have been continually replenished or, as alternative, that there were many sources throughout the area of Peoria deposition. The data presented here indicate that the Peoria of northern Kansas was derived from the Platte and Republican River valleys, largely in Nebraska, and from the Missouri River Valley along the northeastern border of Kansas. The estimated volume of Peoria silt judged to be derived from these valleys, if returned, would more than fill them. Therefore, these outwash-carrying valley sources were not static during the course of loess deposition but were replenished frequently. That late Pleistocene loess deposition in Kansas was pulsatory is clearly demonstrated by the stratigraphic and faunal relationships and correlation with the glacial sections to the north and east. The Peoria is the largest unit of loess in the Great Plains region and is judged to be the product of nearly continuous silt deposition during a significant interval of early Wisconsinan time (Leonard, 1951). The principal evidence that serves to localize source areas includes (1) progressive linear changes in texture, (2) geographic variations in thickness, and (3) minor geographic differences in composition.

Of the three lines of evidence indicative of source areas, change in texture (median particle diameter) is the most impressive. The locations of samples used in seven traverses are shown in figure 2, and the changes in median particle size along these traverse lines are shown in figure 1. The position of major stream valleys is shown on the textural traverses. Five of the seven traverses (*A*, *B*, *C*, *D*, and *F*) originate on the bluff of a major valley—three along the Arikaree River Valley and two along the Platte River Valley in Nebraska.

It is obvious that the coarser particles carried by air currents will settle out nearest the source and that the finer particles will be carried a progressively greater distance. With this criterion, the traverses show the Platte Valley to be a principal source of silt, because the material becomes finer with increased distance, up to several hundred miles, from the Platte. The Arkansas, Missouri, and Republican valleys also are sharply reflected in the traverses, but their effects in Kansas seem to be local. It will be noted that such major valleys as the Smoky Hill, Saline, Solomon, Big Blue, and Little Blue (all lacking late Wisconsinan outwash) show no effect on the textural traverses. The fact that these valleys seem not to have been sources of Peoria loess is confirmed by thickness (fig. 3) and composition data and field examination of deposits along edges of the valleys. In the case of the Arkansas, the small area for which it served as a supplemental loess source is indicated by traverses *A* and *F* (fig. 1). In each, a sharp increase in median diameter is shown immediately adjacent to the val-

ley, but samples south of the valley fall on a projection of the decreasing size curve from the northern source. Griffiths' (1948) measure of sorting, PD_ϕ, which is more sensitive than the quartile deviation, indicates that the best-sorted loess occurs adjacent to the source areas, except for the Arkansas River. This fact, combined with the coarser particle size of the samples near the Arkansas, suggests that this valley was a source of sand rather than of silt.

On theoretical grounds (Krumbein, 1937, p. 585), the decrease in grain size away from a source should be exponential. Therefore, in figure 4 the logarithm of the median diameter was plotted against distance in miles. The straight lines constructed through these points support Krumbein's predictions. These curves (log $Y = a - bX$) differ from that presented by Smith (1942, p. 156) for loess in Illinois, inasmuch as he plotted mean particle size in microns

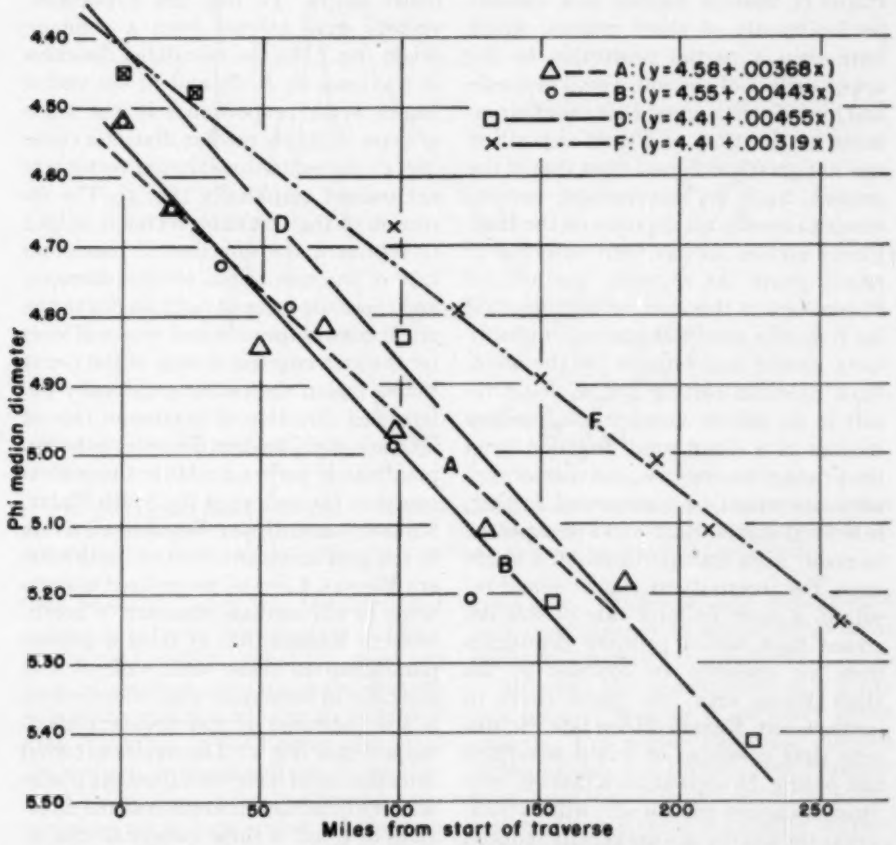


FIG. 4.—Curves showing decrease, away from source, in median particle diameter. Distance in miles is plotted against phi median diameter, and curves were calculated by the least-squares method. Note straight-line relationship and differences in slopes of lines. Slope of line is a function of geographic orientation of traverses as shown in fig. 5.

against the log of the distance ($Y = a - b \log X$). The data from the east bluff of Republican River Valley eastward (fig. 1, *C*) do not fit an exponential curve, for the decrease in median particle size within the first few miles eastward from Republican Valley is too great. It is judged that differences in the environment of deposition between the eastern and western Kansas areas cause the differences in textural curves for the two areas. The floral cover on the High Plains of western Kansas now consists predominantly of short grasses, which form only a partial protection to the upper surface of the soil. Fossil mollusks and buried soils suggest that the environment at the time of Peoria deposition was not greatly different from that of the present. Such an environment permits winds to rework silt deposits on the High Plains surface, as has been observed in recent years. As a result, particles of Peoria loess in this area, which is located far from the principal sources, probably were moved many times by the wind. Such repeated sorting action would result in an infinite number of secondary sources in a down-wind direction from the primary source area, and coarser constituents would be transported farther, in several stages, than could be expected to result from transportation in a single step. The intermittent transit would result in a more uniform rate of size decrease than would primary deposition from air currents. In contrast to the High Plains area, the floral cover in northeastern Kansas during late Pleistocene time consisted of mixed woodland and prairie. In such an area there is very little secondary wind scour; wind velocities at the soil surface are greatly reduced by the protective plant cover; and loess particles, once deposited, could rarely be moved again by the winds—thus the much sharper decline in particle size in the eastern area is explained.

From these data we may conclude that within a relatively homogeneous depositional environment, such as the High Plains, the texture of the loess on the lee side of a linear source should be approximately the same at all points equidistant from the source. Therefore, the direction of maximum rate of increase of phi median diameter (decrease in log size) should be normal to the direction of linear source. To test this hypothesis, vectors were plotted from a common origin (fig. 5) in the calculated direction of traverses *A*, *B*, *D*, and *F;* the vector length being proportional to the angle of slope of the phi median diameter curve (fig. 4). A resultant maximum vector was determined graphically (fig. 5). The direction of the resultant vector is judged to represent the direction of maximum rate of increase of phi median diameter and therefore to be at right angles to the orientation of the principal source of loess for the area covered. A map of the Great Plains region shows the graphically determined direction of maximum rate of increase of phi median diameter to be approximately perpendicular to the general course of the valleys of the South Platte, Arikaree, and Upper Republican rivers in the area north and west of northwestern Kansas. Lines of generalized equivalence of phi median diameter in northwestern Kansas (fig. 2) show a general parallelism to these same valleys. The direction of maximum wind effectiveness is also indicated by this resultant maximum vector (fig. 5). The significant wind direction must have been from the northwestern quadrant, inasmuch as the directions of trend of these valleys all change from northeasterly to easterly or southeasterly within 75 miles of the Colorado state line, whereas the parallelism of the

equal-texture lines is controlled by the northeastward-trending valley segments well past this point.

Textural data indicate that the Platte Valley is the principal source of Peoria loess in north-central and western Kansas; the Arikaree and Republican valleys were important supplemental sources; and the Arkansas Valley contributed importantly, particularly to the coarser size fractions, only in the areas adjacent to it.

region the localities of thickest Peoria loess generally coincide with deposits of coarser texture (fig. 3). Thickness approaching or exceeding 100 feet is found only along the bluffs of the Platte, Arikaree, and Missouri valleys, and deposits approaching 50 feet in thickness occur at many places along the bluffs of the Republican River Valley. Away from these source areas the decrease in thickness is generally proportional to the decrease in

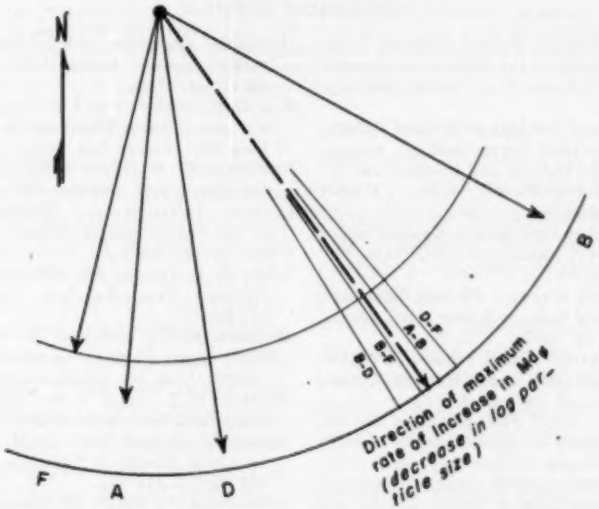


FIG. 5.—Diagram showing direction of maximum rate of increase of phi median size. Orientation of lines indicates average directions of traverses; length of lines indicates relative slope of curve shown in fig. 4; and the resultant maximum length (direction of maximum increase) was determined graphically.

The Missouri Valley was the dominant source in extreme northeastern Kansas, where that valley makes an eastward loop, in a narrow strip along the western side of the valley, and over an extensive area eastward. As the Kansas area is on the western, or windward, side of the Missouri Valley, that valley presumably constitutes a major loess source for the area lying east of it.

Thickness data also furnish strong indications of source areas. In the Kansas

grain size. Thickness data indicate the same relationship to source areas as is indicated by textural variations.

Although the composition of the Peoria loess in Kansas is generally quite uniform, two minor variations furnish strong evidence of source areas. Samples from Doniphan County, extreme northeastern Kansas, contain a significant percentage of detrital dolomite, a mineral not found in samples from other localities in Kansas. As the dolomite occurs

only in samples adjacent to the Missouri River bluffs, it is judged to reflect a difference in source material carried by that stream. Detrital grains of chert are common only in samples from Doniphan, Brown, and Nemaha counties. As the Pennsylvanian bedrock of this area contains chert in significant quantities, in contrast to the bedrock farther west in the state, the presence of chert in the Peoria of northeastern Kansas can be interpreted as due to differences in the source sediment of the Missouri Valley or the incorporation of a small increment of local material in the deposit.

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STRATIGRAPHIC ZONATION OF THE PEORIA LOESS IN KANSAS[1]

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ABSTRACT

An extensive thick sheet of generally calcareous, fossiliferous loess occurs at or near the surface in northern, central, and southwestern Kansas. This loess, the Peoria silt member of the Sanborn formation in Kansas classification, lies above the Sangamon (Loveland) buried soil and includes at the top the Brady buried soil, hitherto of uncertain age. The history of the term "Peorian loess" is reviewed, and the age of the Peoria loess is considered in the light of published reports.

Three zones are recognized in the Peoria loess in Kansas: (1) a basal zone, devoid of molluscan fossils; (2) a lower molluscan faunal zone; and (3) an upper molluscan faunal zone. On stratigraphic and faunal evidence the basal zone is correlated with the Farmdale loess of Illinois, the lower zone with Iowa loess, and the upper zone with Tazewell loess. The Brady soil is shown to have been formed in post-Tazewellian–pre-Caryan (or pre-Mankatoan) time, designated the "Bradyan interglacial interval."

INTRODUCTION

The Peoria loess forms a conspicuous feature of late Pleistocene sedimentation in Kansas, but the problem of the exact age and correlation of this deposit has not previously received attention in published reports dealing with the Pleistocene history of the state. Although it has been generally recognized that the Peoria must represent deposition related to more than one glacial substage, the deposit is a massive silt, lacking obvious indications of subdivisions, and data of sufficient validity to allow detailed correlation of the Peoria complex with named units in the glaciated area of the continent have accumulated slowly. It is the purpose of this paper to present the available evidence pertinent to the age and correlation of the Peoria loess in Kansas and to suggest a probable solution to the problem.

Determination of the age of the Peoria loess is rendered difficult by the inherent complexity of the deposit and by the fact that the term is not everywhere applied to deposits of the same age. Also, it is obscured by the peculiar origin of the term "Peoria" as now applied to a lithologic unit. As Flint (1947, p. 248) has pointed out, little would be lost if the term were allowed to disappear from geologic literature. First recognition of a portion of the lithologic complex now included in "Peoria (or Peorian) loess" was made by Leverett (1898*a*, p. 175) in differentiating what he called Iowan loess, defined as "that sheet of loess which connects at the north with the Iowan till sheet." Leverett (1898*b*, p. 248) was responsible also for introduction of the term "Peorian," which he applied to a presumed interglacial interval between deposition of the Iowan loess and the Shelbyville till. This till was regarded as the earliest glaciation in the Wisconsinan age of the Pleistocene. Leverett based his conception of the Peorian interglacial interval upon a zone of weathering in the upper part of the Iowan loess, and, because the evidence of weathering was particularly noticeable in the loess near Peoria, Illinois, he applied this name to the interval. Alden and Leighton (1917) reviewed the evidence bearing upon the verity of the Iowan glacial stage and, among other things, concluded that the Iowa loess

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was principally post-Iowan in age—that is, the Iowa loess was deposited during Peorian time.

In the first third of the present century considerable attention was given to devising systems of classification of events of the Pleistocene. Kay (1931, p. 449) divided the Pleistocene period (or deposits) into four epochs (or series), for the youngest of which he proposed the name "Eldoran," made up of three parts (called "stages"): Iowan, Peorian, and Wisconsin. He considered the Iowa loess to be Peorian in age. Leighton (1931, p. 45) explained the now common usage of the term "Peorian loess" by stating that "after Alden and Leighton showed that the 'Iowan loess' around the border of the Iowan till sheet had been deposited mostly during early Peorian time . . . usage changed the name Iowan to Peorian." At the same time, Leighton stated his conclusion that the so-called Peorian interval was too short in duration to deserve rank as a stage or substage (then used as time terms in Pleistocene historical geology rather than as rock terms in Pleistocene stratigraphy). Accordingly, he suggested elimination of the term from standard classification. In the meantime, the Wisconsinan till sheets had received detailed study, and Leighton (1933, p. 168) named the Tazewell till ("from Tazewell county, Illinois, where the early Wisconsin deposits are well shown in their relations to the underlying Peorian loess") the Cary till ("from a town in McHenry county, Illinois"), and the Mankato till ("from Mankato, Minnesota"). At this time Leighton proposed a classification of late Pleistocene deposits which placed the Iowa till within the Wisconsin or fourth glacial stage, but so well fixed was the concept of the Iowan glacial interval as an important episode in the Pleistocene that he placed the Iowan at the base of the sequence and called Tazewell, Cary, and Mankato tills "early," "middle," and "late" Wisconsin, respectively. Later that year, Kay and Leighton (1933, p. 673) revised the classification of the Eldoran epoch, dividing it into Wisconsin and Recent stages and reducing Iowan, Tazewell, Cary, and Mankato to substages. Both Leighton (1933) and Kay and Leighton (1933) refused to recognize the Peorian interglacial interval. These substage names are generally accepted today, together with the term "Peoria" (or "Peorian") applied as a lithologic term to Iowan or post-Iowan loess.

With the recognition of the Tazewell and later tills, it became apparent that the Peoria loess splits into two sheets in the area where the Tazewell till is present, one lying above the Tazewell till, although beyond the border of that till the loess is not stratigraphically divisible. It thus becomes obvious not only that the term "Peorian loess" is not in keeping with good practice in geologic nomenclature (because it is a time term used in a lithologic sense) but that, beyond the border of the Tazewell and Iowa tills, the "Peorian loess" is a complex of eolian silts of combined Iowan, Tazewellian, and perhaps later ages (Smith, 1942, p. 142).

In the western Platte Valley, Schutz and Stout (1945, p. 241) recognized a subdivision in the upper part of the Peoria loess consisting of a soil profile overlain by unweathered loess. They named the loess deposit above the soil "Bignell," from a village in Lincoln County, Nebraska, and later (1948, p. 570) applied the name "Brady" to the buried soil. At the same time, probably because the Bignell loess is not everywhere recognizable, they voiced the

opinion that the Bignell is Mankatoan to Recent in age.

In Kansas the Peoria loess lies above the buried soil of Sangamonian age, and it terminates in the Brady soil, which in many places is overlain by the Bignell loess. The problem of bounding the age limits of the Peoria loess resolves itself into the problem of determining the age of the Brady soil, for this profile is more deeply weathered than Recent soil in the same locality and clearly represents a considerable period of relative climatic stability within the Wisconsinan subages. Available data exclude the Iowan-Tazewellian interval from consideration as a possible time for the weathering of the loess to form the Brady soil. Flint (1947, p. 248) points out that the Iowa drift beneath the loess is not leached, and he concludes, accordingly, that "it does not seem likely that the border of the ice sheet receded more than a few hundred miles during the Iowan-Tazewell interval" (p. 249). Similarly, the Caryan-Mankatoan interval does not seem a likely possibility for the time of the formation of Brady soil. Evidence denoting this interval is exposed at Two Creeks, in northern Wisconsin, where lacustrine silts between the Cary and Mankato tills enclose peat, stumps of trees in growth position, and prostrate tree trunks. Nowhere beneath the Two Creeks deposits is the Cary till weathered, and therefore Flint (1947, p. 252) infers that the Two Creeks interval was short. Recently Ruhe (1950, p. 441), from an analysis of drift topographies, concluded that the Tazewellian-Caryan interval constituted the major time break in the Wisconsinan sequence. For these reasons and because Thornbury (1940, p. 471) showed that there is a weathered zone within the upper part of Tazewell till, it seems likely that the Brady soil was formed during a post-Tazewellian–pre-Caryan (or pre-Mankatoan) interval. This is here designated as the "Bradyan interval."

ZONATION OF THE PEORIA LOESS IN KANSAS

In the summer of 1948, investigation of the molluscan fossils in the Peoria loess in Kansas was inaugurated, with a view toward determining whether or not climatic changes concomitant with the presumed several glacial substages represented in this loess were reflected in facies changes in the faunal assemblages and, if so, whether or not faunal zones in the Peoria could be interpreted in terms of subdivisions of the loess within the glaciated region. Toward this end, cubic-foot samples of the loess were taken at short vertical intervals in suitable exposures where the loess could be shown to lie above the Loveland (Sangamon) soil. Shells in the loess samples were recovered by washing the silt through screens. It soon became apparent that the Peoria loess in Kansas could conveniently be divided into zones on evidence obtained from such studies.

This investigation, which has continued to the present time, allows recognition of three well-characterized zones within the body of the Peoria loess. A basal zone of variable thickness is completely devoid of fossil mollusks, but it is in part correlative with the *Citellus* zone of Nebraska classification (Condra, Reed, and Gordon, 1947, p. 30), from which a fossil vertebrate fauna has been obtained. Above this basal zone in suitable exposures, two molluscan faunal zones are found, both lower and upper containing distinctive assemblages of species, which are related by a transitional zone containing elements in common. Each zone is readily recognized in

the field throughout the area of Peoria deposits in Kansas (pl. 1).

THE BASAL ZONE

The lowermost Peoria zone ranges from a few inches to 6 feet or more in thickness. It consists of loess which has been more or less oxidized and to varying degrees leached of free carbonates. The greatest leaching occurs in the basal part of the zone, directly above the Loveland soil "A horizon," where usually no effervescence can be obtained by use of dilute acid and where the silt is generally oxidized to a pinkish-red color. These effects disappear by imperceptible degrees upward into calcareous, unoxidized, and fossiliferous loess. The upper and lower limits of the zone are not precisely definable, for frequently it is impossible to delineate the top of the Loveland soil "A horizon" because the materials in the base of the basal zone (superimposed upon the Loveland soil) may be more or less stained with organic matter, as well as leached and oxidized, and because, at its upper limits, etched and weathered fragments of gastropod shells may appear slightly below the lower faunal zone. Viewed objectively, it is clear that the basal-zone silts were at first deposited so slowly that the loess weathered greatly as it fell. Later, the depositional rate gradually accelerated, until it completely outstripped the rate of weathering of the basal-zone silts, and thereafter relatively unaltered loess accumulated for a considerable time. It may be inferred, though not easily proved, that gastropods were present at the time this silt was deposited. The shells of these snails subsequently were destroyed by weathering processes, because there is no reason to believe that these animals could not have lived under conditions suitable for the vertebrates which left their bones in this basal-zone loess.

THE LOWER MOLLUSCAN FAUNAL ZONE

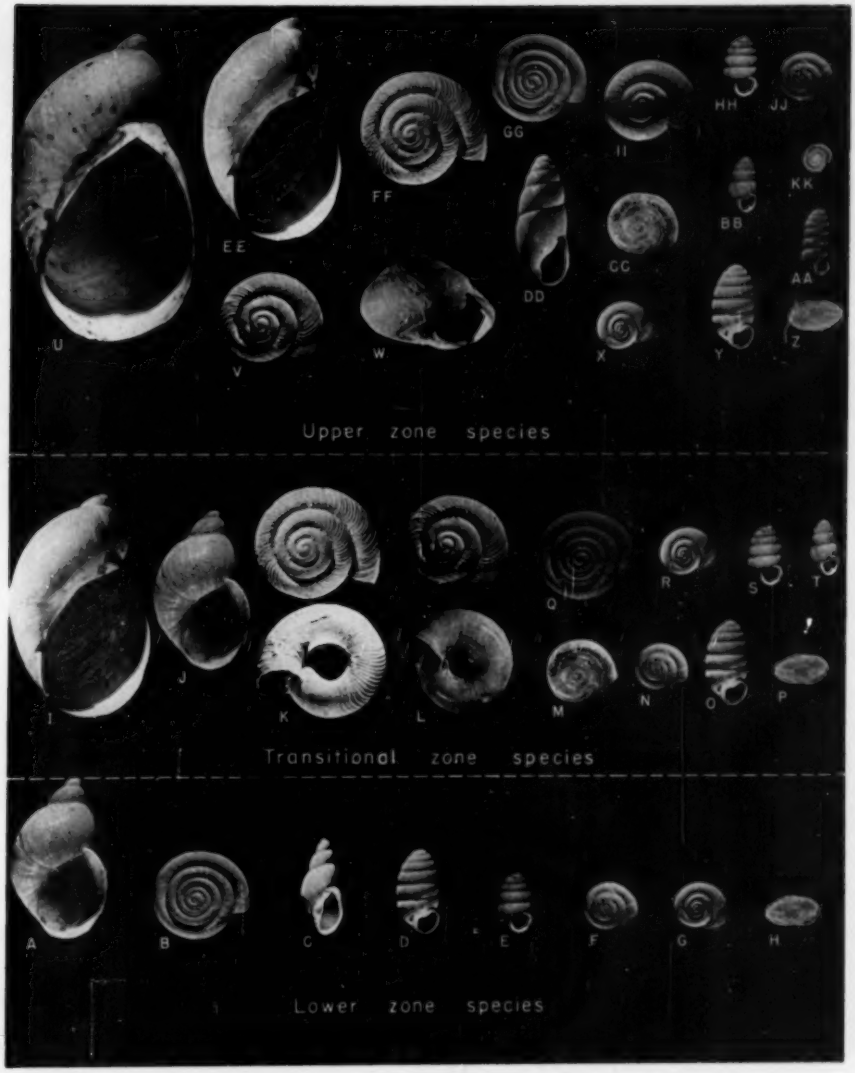
The lower molluscan faunal assemblage contains fourteen species of small terrestrial gastropods, two of which are restricted to this zone. The others appear also in the upper faunal assemblage. The occurrence of restricted species, though few, associated with the universal absence of no less than fourteen species which have been found only in the upper faunal assemblage, makes the lower fauna distinctive. It is readily discernible at outcrops of the loess.

Among the lower-zone species, *Vallonia gracilicosta*, *Pupilla muscorum*, *P. blandi*, *Vertigo gouldi paradoxa*, and *Helicodiscus singleyanus* are extinct in Kansas. Most of the remaining species are restricted to favorable situations in the eastern part of the state. A few, including *Succinea avara*, *Hawaiia miniscula*, and *Deroceras laeve*, are living over most of the area considered in this report, especially in locally favorable habitats.

The most widespread species of the lower faunal zone are *V. gracilicosta*, *S.*

PLATE 1

Some of the species of most common occurrence in the faunal zones recognized in the Peoria loess of Kansas. All figures are approximately 3× natural size. *A, J, Succinea avara. AA, Columella alticola. B, Q, GG, Helicodiscus parallelus. C, Lymnaea parva. D, O, Y, Pupilla muscorum. E, S, II, Pupilla blandi. F, N, JJ, Hawaiia minuscula. FF, K* (dorsal and ventral views), *Discus cronkhitei. G, R, X, Vallonia gracilicosta. H, P, Z, Deroceras laeve. I, EE, Succinea grosvenori. KK, Striatura milium. M, CC, Helicodiscus singleyanus. T, BB, Vertigo modesta. V, L* (dorsal and ventral views), *Discus shimeki. U, Succinea ovalis. W, Hendersonia occulta.*



Common faunal species of the Kansas Peoria loess

avara, *P. muscorum*, *P. blandi*, and *H. minuscula*, which occur at almost every locality reported, including those in southwestern Kansas, where conditions, then as now, seem to have been generally unfavorable for terrestrial gastropods. Because three of these species are extinct in the state, the lower faunal assemblage is readily recognizable, even in southwestern Kansas. Here the loess is thin, fully half its thickness being involved in the topsoil profile, from which any shells formerly present have been removed by weathering. One species, *Lymnaea parva*, sometimes considered an aquatic animal, although actually littoral in habit, may occur in the loess at the sites of ponds on former loess surfaces. *Carychium exiguum* belongs in this same category.

Above the basal zone the lower faunal assemblage increases progressively upward, both in number of species and in individual shells. For example, at locality 18 (Decatur County), the lower 5 feet of a 25-foot exposure of Peoria loess contains no shells (basal zone); 7 feet above the base, 139 shells grouped in 4 species were recovered from a cubic foot of silt; while 13 feet above the base a cubic foot of loess yielded 753 shells, assigned to 6 species. At this same locality the transitional-zone assemblage consists of 11 species and a total of 5,079 shells per cubic foot; the upper faunal zone comprises 14 species and a population of 1,578 shells per cubic foot of loess.

THE TRANSITIONAL ZONE

Between the upper and lower faunal zones in the Peoria loess there is a transitional faunal zone, which bears elements of both the lower and the upper faunal assemblages. Not regarded here as a significant subdivision in the Peoria loess, this zone nevertheless serves to verify the absence of any important change in the rate of deposition during the accumulation of the main body of the Peoria and confirms the absence of any demonstrable lithologic unconformity in this complex silt. Conditions seem to have been favorable for mollusks during this interval, and it is possible that depositional rates slowed somewhat, if one may judge from the fact that the highest population density occurs in the transitional zone. It is, however, obvious that depositional rates did not at any time become sufficiently retarded to allow weathering processes to remove much free carbonate from the loess or to destroy any appreciable proportion of the shells.

Discus cronkhitei and *D. shimeki* are the first of the upper-zone species to appear in the transitional zone, generally in association with *S. avara*, which is otherwise restricted to the lower faunal assemblage. *Succinea grosvenori* and *S. ovalis* appear higher in the transitional zone, after the disappearance of *S. avara*. Other upper-zone species which may appear in the transitional zone are less widespread, and their occurrence in this zone seems to reflect local conditions rather than general climatic change.

THE UPPER MOLLUSCAN FAUNAL ZONE

The upper molluscan faunal zone (fig. 1) comprises 26 species, including 14 species which do not occur in the lower faunal assemblage. Four species—*Columella alticola*, *Striatura milium*, *Vertigo coloradensis*, and *D. shemeki*—do not live within the Great Plains region, nor are any of these species known from any Pleistocene or older horizon in Kansas. *Discus cronkhitei*, *P. muscorum*, *P. blandi*, *V. modesta*, and *Helicodiscus singleyanus* are likewise extinct in the Great Plains, but they are known from earlier Pleistocene deposits in Kansas

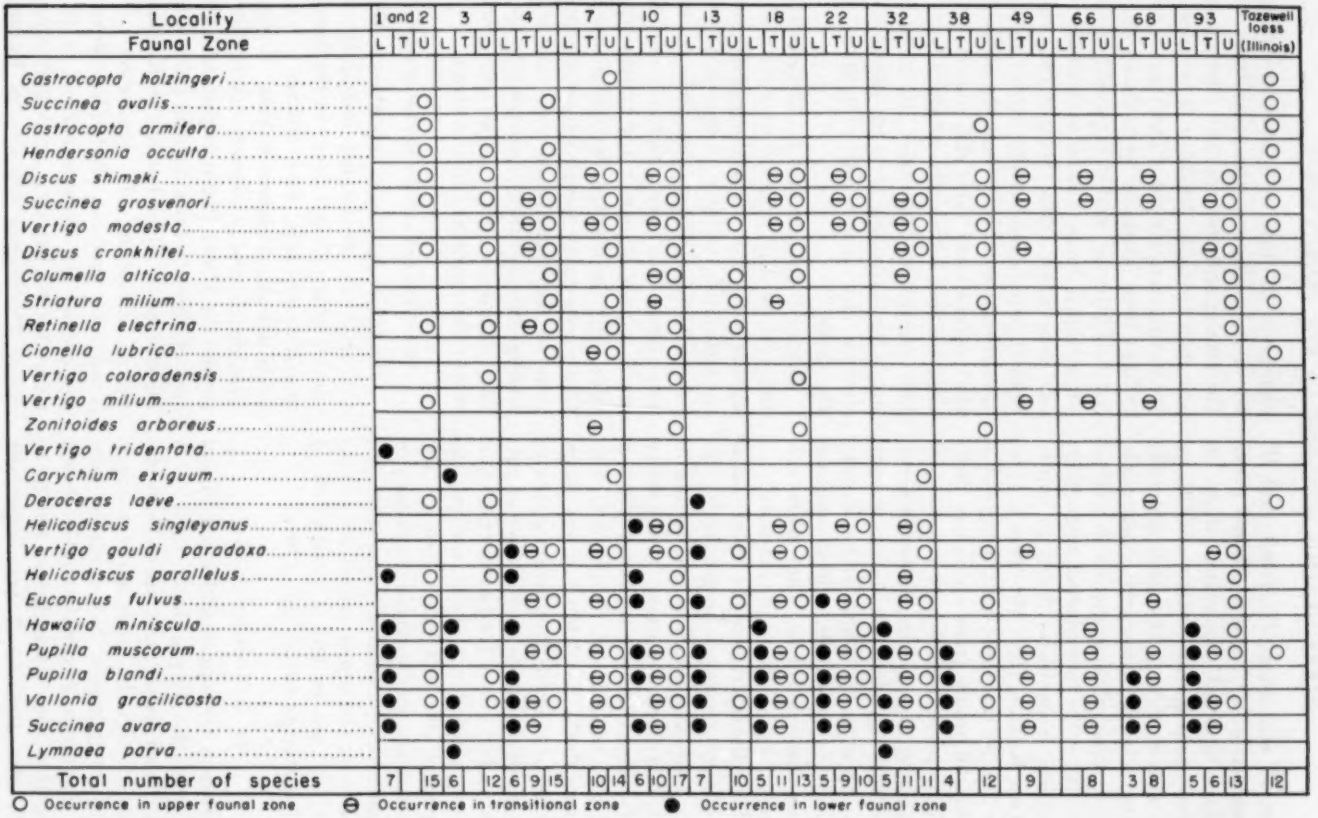


FIG 1.—Zonal distribution of fossil mollusks in Peoria loess at selected localities in Kansas (1–68), Nebraska (93, Peoria loess near Bignell), and Illinois (Tazewell loess near Creve Coeur). Location of localities 1–68 is shown by Frye and Leonard (1951, fig. 1).

(Frye and Leonard, 1951, fig. 2), the last only from the lower faunal zone of the Peoria. In all, 11 of the 26 species in the upper faunal assemblage are extinct in the state, making this zone easily recognizable in the field. The remaining species are not generally present in the area under consideration, being restricted to local situations, particularly in northeastern Kansas. As pointed out above, conditions were more favorable than now for mollusks during the time of deposition of the loess containing the transitional and upper faunal assemblages, for shells are present in great numbers, commonly exceeding 5,000 shells per cubic foot of loess. The greater variety of species, compared with the living fauna in the area, and the presumed ecological requirements of the species now extinct in the state confirm this conclusion.

ECOLOGICAL IMPLICATIONS OF THE FAUNAL ZONES

It is not possible to interpret very accurately the ecological conditions associated with Peorian deposition because ecological factors which limit the distribution of terrestrial gastropods are inadequately known. Specifically, it is not known why the Peorian species now extinct in Kansas and in the Great Plains generally no longer live there. It is clear, however, that ecological conditions during Peorian time were relatively favorable to terrestrial gastropods, inasmuch as the fossil faunas, even in the lower faunal zone, are more varied than Recent faunas in the same areas. It is likely that population density was also greater, judging from the large numbers of shells recoverable from a cubic unit of loess, but the rate of deposition is a factor which cannot be determined by any known means, so the actual number of shells per cubic unit may be misleading.

A reasonable amount of rainfall and a floral cover at least as dense as that now prevailing may be inferred over the area in Kansas involved in Peorian deposition, as terrestrial gastropods are active and can reproduce only during intervals when the soil and overlying organic matter are moist. For two reasons it may be judged that temperatures were slightly lower than at present: (1) the Peorian species now extinct in the state presently live at higher altitudes or latitudes, and (2) there is no record in the loess of *Bulimulus dealbatus*, a gastropod of southern affinities which has reached the northern border of Kansas in Recent times. This species is adapted to arid conditions, which may indicate that rainfall was greater during the time of Peoria deposition than now.

Although it is true that Peorian species now extinct in the state live at higher altitudes or latitudes than they did when they populated Kansas, the exact factors which brought about their extinction in this region and presently limit them in distribution are not known. Extreme minimal temperatures on the Great Plains are not greatly at variance with those prevailing in the regions now occupied by these locally extinct species, nor is there any appreciable difference in average annual rainfall. A possible limiting factor which may account for the extinction in Kansas of the Peorian species is the occurrence in the Great Plains of extremes of high temperature and severe aridity. Despite little direct evidence, it may be inferred that these phenomena are responsible for the extinction of species which are abundant in the Peoria loess. It seems reasonable to assume that the presence of continental and Rocky Mountain ice sheets would

exert a stabilizing influence on temperatures and rainfall, without necessarily lowering average annual temperatures in the Great Plains appreciably.

It is equally clear that ecological conditions changed during the whole interval in which the Peoria loess in Kansas was being deposited, because the upper-zone fauna contains so many new elements and because it *seems* that population density increased during this interval, perhaps a great deal. In view of these observations, attempts were made to find suitable loess faunas in the glaciated region where the loess could be associated positively with a single glacial interval. I am indebted to Dr. H. B. Willman, of the Illinois Geological Survey, for the opportunity to collect a molluscan fauna from undoubted Tazewell loess, in an exposure near Creve Coeur, Illinois. This faunal assemblage is an exact correlative of the upper-zone fauna in the Peoria in Kansas.

CORRELATIONS OF THE PEORIA LOESS

It now seems possible to limit the age of the Peoria loess in Kansas and to correlate the zones within it with recognized Pleistocene units elsewhere. In Nebraska, Iowa, and Illinois, as well as in Kansas, the Peoria complex almost invariably has at its base a somewhat weathered zone, which generally grades upward into calcareous, fossiliferous loess but which at a few places in Nebraska, Iowa, and Illinois is terminated above by a weakly weathered zone. This loess, wherever I have seen it over these four states, is most greatly leached and oxidized in its basal part, which has a pinkish color. It has been called the "Farmdale loess" by Leighton and Willman (1949, p. 1904) from the Farm Creek section in central Illinois. At this place this unit has considerable organic matter in its upper few inches, indicating a slowing of the rate of deposition.

The best explanation for the Farmdale interval seems to lie in the assumption that the Iowa ice advanced and then receded slightly, producing outwash valley trains from which silts were air-borne to deposit slowly as the basal-zone (Farmdale) loess, before advancing to its final stage, destroying in the process the evidence of its earlier stand. Whatever may be the true explanation for the origin of the basal-zone loess, the identical stratigraphic position, nature and degree of weathering, and similar lithology throughout a wide region conclusively indicate that the basal zone of the Peoria complex in Kansas is equivalent to the Farmdale loess of Illinois classification.

The slightly weathered zone in the upper part of the Farmdale loess in Iowa and Illinois, the presence of organic matter in the upper few inches of the Farmdale loess at the Farm Creek section, and the lack of these features in Kansas are susceptible of more than one explanation. The Peoria loess in Kansas clearly has two general sources; that in eastern Kansas is derived from Missouri River valley trains, whereas the loess in central, western, and southwestern Kansas originated from such valleys as that of the Platte, Arikaree, and Republican rivers (Swineford and Frye, 1951), which, during late Pleistocene glacial episodes, were under the influence of Rocky Mountain, rather than continental, glaciation. Peoria loess in northeastern Kansas is not generally exposed at its base, but in the few available exposures no incipient soil has been noted in the top of the basal zone. This may be due to purely local conditions of drainage or erosion, although no erosional unconformities have been recognized. The loess in the remainder of the state fell in an

area of presumed lesser rainfall, which might account for the absence of a weathered zone in the top of the basal zone. It is also possible that the time between the inferred initial advance of the Rocky Mountain glaciers of Iowan age and their readvance after a slight retreat was less than that of continental ice generally synchronous with it.

Because the upper faunal-zone assemblage is identical with that found in known Tazewell loess and inasmuch as the lower and upper faunal zones are separated neither by a faunal hiatus nor by a lithologic or erosional unconformity, even in the absence of a molluscan fauna from restricted Iowa loess, it seems appropriate to correlate the loess containing the lower-zone faunal assemblage with Iowa loess (Iowan and post-Iowan in age) of Iowa and Illinois and the loess containing the upper faunal assemblage in the Peoria of Kansas with Tazewell loess (Tazewellian and post-Tazewellian in age) of Illinois.

AREAL EXTENT OF IOWA AND TAZEWELL LOESSES IN KANSAS

The lower faunal zone has been traced from northeastern Kansas through central, western, and southwestern parts of the state and is present in nearly all localities listed by Frye and Leonard (1951, fig. 1). The same faunal assemblage has been obtained in northwestern Oklahoma in Dewey County, in northern Texas in Sherman County, and in Frontier, Lincoln, and other counties in Nebraska. Because it is judged that the loess in which these faunal assemblages occur is largely retreatal Iowan in age, it may be safely assumed that Iowan valley trains were the source of loess over this region.

The upper faunal assemblage is much more restricted in its distribution in Kansas but is known from southern and central Nebraska, western Iowa, and central Illinois. This assemblage has been traced across the northern border of Kansas (except for Cheyenne County in the extreme northwestern part of the state) and southward to Rush, Lane, Gove, and Logan counties. It is concluded from the faunal evidence that these localities represent the approximate southern border of the greatest extent of the influence of Tazewellian ice upon loess deposition in Kansas. However, inasmuch as the Peoria loess of southwestern Kansas is thin, nearly half its thickness being involved in the Recent soil profile, it is possible that loess blown from Tazewell valley trains was carried even farther south, evidence having been lost by weathering of the shells.

It becomes more clear that the Brady soil was formed during a post-Tazewellian–pre-Caryan (or pre-Mankatoan) interglacial interval, since this fossil soil is developed in the upper part of loess shown to be Tazewellian in age. The name "Bradyan" is proposed for this interglacial substage.

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LOESS AND QUATERNARY GEOLOGY OF THE LOWER MISSISSIPPI VALLEY[1]

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ABSTRACT

A wealth of evidence demonstrates the Pleistocene age of the four depositional terraces which border the lower Mississippi alluvial valley and form the broad coastwise plains of Louisiana and Mississippi. This evidence also conclusively proves that the Mississippi Valley alluvium is of Recent age and fills a deeply entrenched valley system eroded during the late Wisconsin glacial stage.

Both the Recent age and the mineralogy of the alluvium make it an impossible source for the several loess "formations" as described in the region by M. M. Leighton and H. B. Willman. Furthermore, the age of their loess "formations" does not conform with the age of the Pleistocene terrace deposits which the loess mantles.

The distribution of lower-valley loess, its topographic expression, the absence of associated sand dunes and fossil trees, and the presence of colluvial features make it difficult to accept the hypothesis of eolian origin for loess in this region. The loessification theory held by R. J. Russell is the only hypothesis yet proposed that is in accord with known facts.

INTRODUCTION

Attention has again been focused upon the important matter of the age and origin of the lower Mississippi Valley loess by the recent article, "Loess Formations of the Mississippi Valley," by M. M. Leighton and H. B. Willman (1950). These writers have used a stratigraphic approach in an effort to correlate the lower-valley loess with loess formations in the upper valley. Although the approach has been used previously and similar correlations have been made by Wascher, Humbert, and Cady (1948), the article warrants attention because it is the first such attempt in several decades to carry the weight of authority. Not only was it written by men long recognized as leaders in the field of Pleistocene geology, but it is the outgrowth of studies which culminated in a 2 weeks' field conference held in June, 1949. This conference, designed to harmonize divergent views concerning the geology of the upper- and lower-valley regions, started at Iowa City, Iowa, and was concluded near Natchez, Mississippi. It was sponsored by state geologists of the region and was attended by many Pleistocene specialists.

Leighton and Willman believe the loess to be stratigraphically continuous throughout the Mississippi Valley region. Therefore, they contend that, inasmuch as the upper-valley deposits are considered to be eolian, those in the lower valley must have the same origin. They hold that the lower-valley loess was deposited during glacial stages and was derived from those courses of the river which deposited silt and sand. At several localities they found three superimposed layers of loess which, on the basis of weathering, they correlated with loess formations of Illinois (fig. 1). In addition, they think that a fourth and older deposit is present at several places.

Such findings are in marked contrast to those of Russell (1944), who believes that the lower-valley loess developed through a process of loessification, i.e., that materials which are weathered from backswamp deposits in Pleistocene terraces accumulate on slopes through colluvial action and change to loess as cal-

[1] Manuscript received January 27, 1951.

cium carbonate is added by ground water. Because the process of loessification cannot be precisely dated, Russell feels that loess can have no stratigraphic significance other than its Quaternary age.

In fitting their findings on loess into the framework of lower-valley stratigraphy, Leighton and Willman challenged the present writer's interpretation of the late Cenozoic geology of the region, without adequate consideration of the data upon which it was based. Although their exact viewpoint on stratigraphy and the late Cenozoic history of the lower-valley region is somewhat difficult to obtain from the article under discussion, the interested reader will be able to find it precisely stated in several recent articles and abstracts (Leighton and Willman, 1949*a; b*, p. 5*a; c*).

Inasmuch as a knowledge of the basic facts of the Quaternary geology of the lower-valley region is prerequisite to an appraisal of the current controversy concerning the age and origin of loess, the data and their interpretation are herewith presented.

LOWER VALLEY		UPPER VALLEY			
FORMATIONS	EVENTS	STAGES		SUBSTAGES	MATERIALS
(Fisk, 1938)		(Kay, 1944)		(Leighton & Willman, 1950)	
Recent	Alluviation	Recent			
	Valley Cutting	Wisconsin	Wisconsin	Mankato	Loess* / Till
				Cary	Loess* / Till
Prairie	All.	Peorian		Tazewell	Loess* / Till
	V.C.	Iowan		Iowan	Loess* / Till
				Farmdale (Pro-Wis.)	Loess**
Montgomery	All.	Sangamon			Gumbotil
	V.C.	Illinoian		Buffalo Hart Jacksonville Payson	Till
				Loveland (Pro-Ill.)	Loess**
Bentley	All.	Yarmouth			Gumbotil
					Till
	V.C.	Kansan		Pro-Kansan?**	Loess**
Williana	All.	Aftonian			Gumbotil
	V.C.	Nebraskan			Till

Peorian Loess**

* Peorian Loess where not differentiated

** Recognized in lower valley by Leighton & Willman, 1950.

FIG. 1.—Quaternary stratigraphy of the upper and lower Mississippi Valley regions

Summary of Lower Mississippi Valley Quaternary Geology

The physiography of the central Gulf coastal plain is dominated by a broad belt of coastal marshes, by the wide alluvial plain of the Mississippi River, and by a series of four steplike coastwise depositional terrace plains which rise northward in Louisiana and Mississippi (the names of these terraces are given in fig. 1, and their distribution in southern Louisiana and Mississippi is shown in fig. 2). Each coastwise plain occurs as a "paired" terrace on opposite sides of the Mississippi Valley and provides evidence as to the amount of regional uplift which has taken place subsequent to its formation. The coastal-plain terraces are ancient deltaic plains of the Mississippi River and coastal marshlands. Their succession marks the progradation of the Gulf shore which has taken place contemporaneously with uplift of the lower-valley region. Each of the coastwise plains merges upvalley into an equivalent depositional terrace bordering the Mississippi alluvial plain.

Data from more than twenty-five thousand borings and from exposures show that a thick mass of alluvium underlies the alluvial plain and that similar, independent masses underlie each terrace. As the deposits are mappable, they were given formational rank by the writer (Fisk, 1940, p. 175). The sediments associated with the stream terraces are definitely fluviatile, but borings indicate that the coastwise plains and the coastal marshlands are underlain by thick deltaic masses in which marine layers are interbedded with fluviatile deposits. The valley alluvium and each of the terrace formations are of irregular thickness, and each rests upon an erosional surface which was deeply incised across the central Gulf coastal plain. The sediments of each formation grade upward from a coarse substratum of gravel-bearing sands into a top stratum of sands, silts, and clays. The topographic positions and the relationship of the two higher terrace formations and the alluvium of the Mississippi Valley to underlying erosional surfaces are indicated on the cross section (fig. 3).

The terrace surfaces bear traces of meandering streams comparable in size to the modern Mississippi, providing evidence that conditions during the construction of the terrace plains were much as they are today. Ancient meander patterns, reconstructed from drainage lines and from soils and vegetation patterns present on aerial photographs, have been previously shown on photographs and maps published by the writer (Fisk, 1944, figs. 72, 79, 80). The best-preserved relict meanders are on the little-eroded coastwise plain of the Prairie terrace in southern Louisiana, where meander belts of both the late Pleistocene Mississippi and Red rivers are clearly visible; reconstructions of these features are shown in figure 2. Contrary to the statement made with reference to reconstruction of similar features on older terraces (Leighton and Willman, 1950, p. 621), these meander belts are not reflections of pre-loess drainage lines through the loess but are well-defined surface features; wide lowlands mark relict channels, and gentle slopes away from the channels mark the positions of natural levees. The local relief on well-preserved Pleistocene surfaces is similar both in amount and in kind to that of the Recent floodplain.

AGE OF TERRACE FORMATIONS

When the stratigraphic and physiographic relationships are interpreted in

FIG. 2.—Physiographic features of southern Louisiana

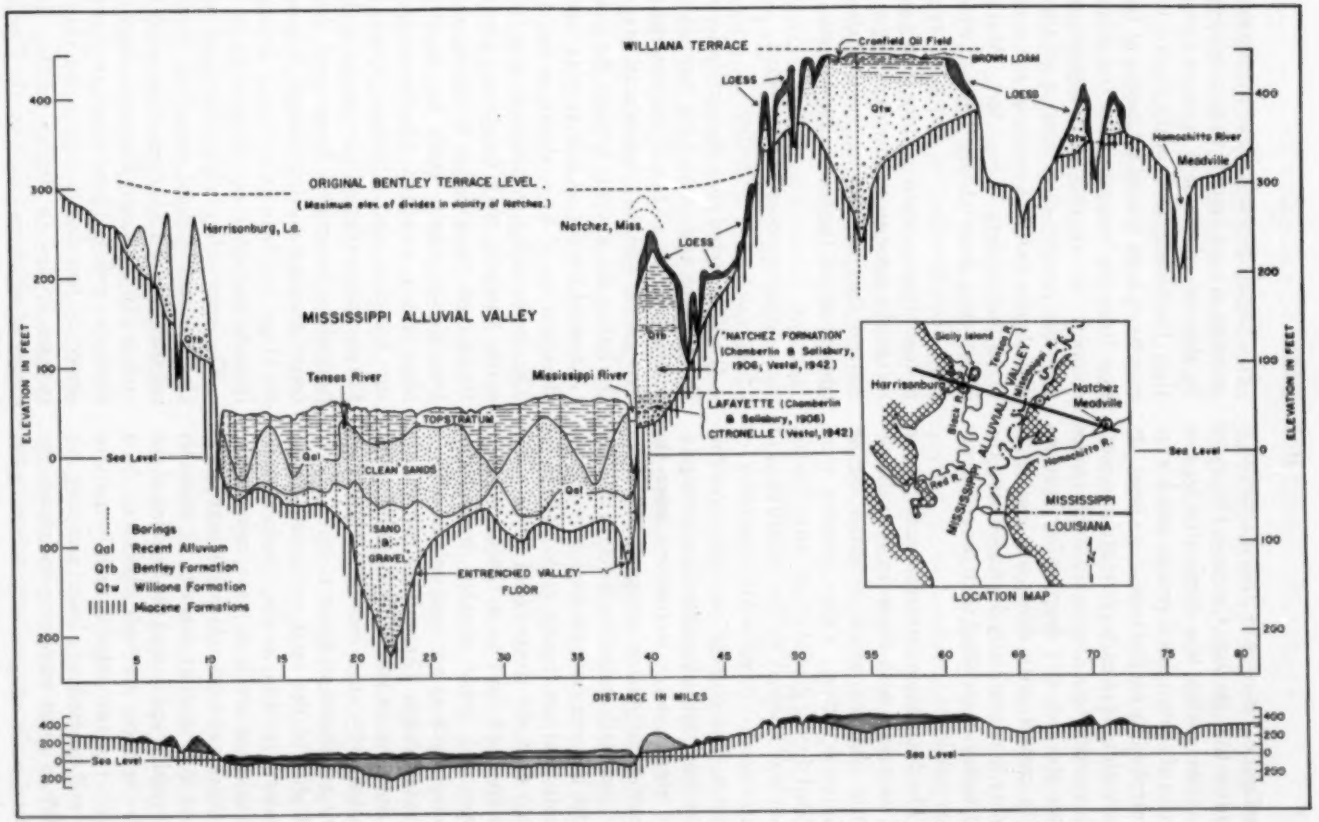

FIG. 3.—Cross section of lower Mississippi Valley region in the vicinity of Natchez, Mississippi

the light of changes in stream activity, it is plain that the late Cenozoic history of the lower valley was dominated by five cycles of stream development which were controlled by oscillations in sea level. At the beginning of each cycle, streams were rejuvenated by a drop in sea level of more than 400 feet and deep valley systems were carved; with rising sea level, gradients were gradually decreased, and the cut valleys were filled with progressively finer particles.

The fluctuations in stream activity can be correlated with the series of cyclic climatic changes which occurred during Quaternary time. Valley systems were eroded during glacial stages when the sea level was lowered by the withdrawal of water needed to maintain expanding ice masses. The entrenched valley systems were filled with alluvium during interglacial stages, when meltwaters from the diminishing glaciers returned to the sea and gradually raised its level. It was not until the sea had reached a relatively stable position during each interglacial stage that the rivers could adjust their volume and load to a constant valley slope and could develop meandering courses such as are seen on the modern alluvial plains. The writer's correlations of the terraces and their underlying formations with other events of Pleistocene time are shown in figure 1.

Uplift of the land continued during Quaternary time as the glacio-eustatic shifts in sea level took place; consequently, the inland terrace formations are separated in a steplike manner. Erosion accompanied uplift, and the terrains of the older terraces are well dissected. As a result, the two older terrace formations are well exposed over broad areas (fig. 3).

While uplift was going on inland, the coastal plain was gradually subsiding beneath a great mass of deltaic sediment (Fisk, 1939, p. 192); thus the seaward portions of each terrace were overlapped by deposits of the next younger formation. The basal gravels of the highest terrace have been traced by means of borings from the outcrop into the deltaic complex, where they are encountered near the present shore line in southern Terrebonne Parish, Louisiana, at a depth of approximately 3,000 feet. Marine Pleistocene invertebrate faunas, which have been described by Lowman (1949, p. 1990) from a depth of more than 3,500 feet in the same area, provide strong evidence of the Quaternary age of the deltaic equivalents of the terrace formations.

INTERPRETATIONS OF PLEISTOCENE CHRONOLOGY

It will be noted that the correlation of the terraces and underlying formations shown in figure 1 fits well the concept of five glacial and five interglacial stages (including the Recent) held by the Iowa Geological Survey as late as 1944 (note classification of Kay, shown in fig. 1). The lower-valley correlation does not agree, however, with Leighton's (1931) Pleistocene classification involving only four glacial stages, which has recently been widely accepted. He harmonized the American and European views of Pleistocene stratigraphy by grouping the previously recognized Iowan and Wisconsin glacial stages into a single Wisconsin stage, which he later (1933) divided into the four substages shown in figure 1.

The lower valley holds evidence of two glacial stages of equal magnitude covering that part of Pleistocene time which is included in the single Wisconsin stage in the upper valley. These stages are represented by two deeply entrenched valley systems eroded to similar depths, the earlier of which was filled by the Prairie formation and the later by the Recent

alluvium. Three older entrenched valley systems, also of almost equal magnitude, complete the Pleistocene chronology. They correspond to the Nebraskan, Kansan, and Illinoian glacial stages and exist beneath the Williana, Bentley, and Montgomery formations.

Because differing views as to late Pleistocene chronology affect the interpretation of the origin, age, and source areas of the loess and because the broad aspects of the Quaternary history are so clearly shown in the lower-valley region, the validity of a fourfold classification needs examination. No criticism is directed at either the mapping of features in the deglaciated terrain of Illinois or the stratigraphic sequence which has been developed there. The issue is confined wholly to the advisability of interpreting events which are portrayed by substages in Illinois as having world-wide significance and using them as a basis for interregional correlations. It seems entirely possible that the broader aspects of late Pleistocene chronology could have been obscured in Illinois by details of erosion and deposition caused by local shifts in the lobate front of the ice sheet. The events of glaciation revealed by detailed mapping of the Wisconsin drift sheets in Illinois are known in few other regions; before a fourfold chronology can be proved universally applicable, equally detailed studies in other deglaciated areas throughout the world are needed.

To the knowledge of the writer, Russell (1940, p. 1222; 1941, p. 82) was the first to suggest that nonglaciated regions might afford the best areas from which to interpret world-wide effects of glaciation. He was no doubt influenced by the fact that only major events of a glacial stage are reflected in such places. The upper and lower limits of an alluvial mass near the Gulf shore, indicating the maximum and minimum of a cycle of sea-level change, record only extremes in the total volume of ice stored on the continents. A minor advance or retreat of the ice, during either a waxing or a waning of the ice sheets, would have taken place at a time when the sea level was low. The alluvial deposits in the lower valley which record such minor movements are buried by materials laid down as the sea level subsequently rose. Although such details are present in each alluvial formation and probably could be interpreted, they in no way complicate the interpretation of the major events of Quaternary time.

DISTRIBUTION AND STRUCTURE OF TERRACES

There can be little question that Leighton and Willman doubt the existence of the Pleistocene depositional terraces as mapped by the writer, for they consistently refer to them as "terraces" and make unsupported statements to the effect that, in mapping them, he confused silts with backswamp deposits (see Leighton and Willman, 1950, p. 621). On this basis they assert that he incorrectly correlated fluviatile terraces with older alluvial surfaces west of the Mississippi Valley in central Louisiana and developed an incorrect plane of reference for diastrophic interpretations (Leighton and Willman, 1950, p. 622). In view of this criticism, it seems necessary to point out that the mapping of the terraces was not based solely upon their relative elevation and distribution; a great deal of information concerning the underlying formations was gained from exposures and from samples and logs of many borings. Furthermore, the correlations of the coastal-plain surfaces in southern Louisiana and Mississippi (fig. 2) with their equivalent fluviatile terraces upstream

were established by direct tracing in the field.

A generalized map showing the distribution of the terrace surfaces in the lower Mississippi Valley has been published by the Mississippi River Commission (Fisk and others, 1949, pl. 4). Apparently no consideration was given to the data included in this report or to those of previous publications by the writer (Fisk, 1938, 1939, 1940, 1944, 1947). These reports not only discuss the basis for mapping the terraces but also contain many cross sections and diagrams showing their nature and distribution, together with their underlying formations.

The fluviatile terraces are present on both sides of the Mississippi alluvial valley. On the eastern side they make up a belt averaging less than 25 miles wide, generally coincident with the Bluff Hills, and extend from southern Mississippi as far north as the Cache River lowland in southern Illinois. They also occur on Crowley's Ridge, the narrow 200-mile-long upland remnant in the northern part of the valley between Cape Girardeau, Missouri, and Helena, Arkansas. The youngest terrace is present at places on Macon Ridge, which extends down the center of the alluvial valley for 100 miles from Eudora, Arkansas, to Sicily Island, Louisiana. The terrace formations have been mapped in northeastern Louisiana and southeastern Arkansas, where they border the western side of the valley, but only incompletely north of Little Rock, Arkansas.

In all field work done by the writer the presence of loess was recognized but did not enter into the determination of any terrace level. Only the maximum elevations of the little-eroded divide areas were used in computing the original elevations of the depositional terrace plains upon which structural interpretations were based (Fisk, 1939, 1940, 1944). Despite statements to the contrary (Leighton and Willman, 1950, p. 622), at no place in the lower Mississippi Valley has loess associated with a given terrace formation been found at a higher elevation than the original surface of the depositional plain. Typical loess occurrences in the eroded terrain of the two higher terrace formations in the vicinity of Natchez, Mississippi, are shown in figure 3.

There is no evidence in the distribution of the depositional terraces that Pleistocene alluviation was more widespread throughout the region at any stage than it is on the modern plain, though the writer has been interpreted to the contrary (Leighton and Willman, 1950, p. 622). As a result of regional uplift and progradation of the Gulf shore throughout the Quaternary, fluviatile plains have developed within the terrain of older coastwise plains (fig. 2). The deltaic plain of the highest terrace, the Williana, is outlined in the "delta" shape of its outcrop (Fisk and others, 1949, pl. 4). The apex of the "delta" lies near Jackson, Mississippi, more than 150 miles inland. Post-Williana uplift and outbuilding of the shore have permitted the establishment of the relatively narrow fluviatile plain of the Mississippi Valley within a terrain of broad coastwise plains. That alluviation is no less widespread today than in Williana time is shown by the way in which the modern alluvial plain broadens into a coastwise deltaic plain about as wide as that of the Williana.

It is not clear whether Leighton and Willman (1950, p. 622) believe post-Aftonian uplift to be excessive or nonexistent. The uplift certainly exists and was found by this writer to be more than 400 feet in the latitude of Natchez, Mississippi, rather than 300 feet. That they also

recognized a considerable amount of Quaternary uplift can be assumed from the following quotation (Leighton and Willman (1950, p. 614):

> The authors, after careful field study, have reached the opinion that the Natchez formation is a Pleistocene valley-train deposit, as interpreted by Chamberlin (1896), Chamberlin and Salisbury (1907, pp. 386–388), and Vestal (1942), and is separate from, and younger than, the Lafayette-type gravels, contrary to Fisk (1938, pp. 152–154).

Vestal (1942, p. 46) recorded the Natchez formation at the head of Solitary Valley Creek, near Natchez, Mississippi. This locality is shown on the Natchez quadrangle (U.S. Geological Survey) to have an elevation of more than 300 feet. The elevations show that the Natchez formation has been uplifted approximately 250 feet with respect to the alluvial plain. Leighton and Willman did not mention the evidence from their field study which permitted them to call these beds a Pleistocene valley train, or the particular glacial stage during which the Natchez formation was deposited. However, according to the writer, the "Natchez formation" is but the sand facies of the Bentley formation (fig. 3), and the uplift has taken place since Yarmouth time.

MINERALOGY OF THE TERRACE FORMATIONS

The distribution of terraces in the eastern uplands is of particular interest because, south of the Ohio River, the belt in which they are found is bordered on the east by higher hills of Tertiary sediments. All the terraces can be traced directly into the upper Mississippi and Ohio River valleys, thus proving that the sediments of all the terrace formations came from the same general source area. This fact limits the interpretation (Leighton and Willman, 1950, p. 621) that heavy minerals of the basal facies of the older terrace formations, the so-called "Citronelle" gravels, are like those of the Tertiary and distinct from those of the younger formations. No age distinctions need to be invoked in such findings because heavy minerals of the high-terrace gravels on the eastern side of the valley were probably in large part locally derived from the Tertiary formations and were mixed with those carried by the master-stream. During the first glacial stage, when lowered sea level brought about pronounced stream entrenchment, the Tertiary beds were deeply eroded and were the source of great quantities of bed-load materials supplied by local streams to the master-river.

According to Grim (1936) and Bornhauser (1940), Tertiary beds in the lower-valley region contain a greater percentage of kyanite, staurolite, and zircon than of other heavy minerals, as do the "Citronelle" gravels (Leighton and Willman, 1950, p. 621). Also occurring in the gravels, but in varying minor percentages (Doeglas, 1949, p. 116; Leighton and Willman, 1949*b*, p. 79), are the heavy minerals hornblende, garnet, and epidote, which characterize the bed load of the modern Mississippi River (R. D. Russell, 1937). The presence of these minerals provides evidence of Mississippi River activity in the region during the early glacial stage and supplies an index as to the relative amount of heavy minerals brought into the valley by local tributaries and by the Mississippi. While the sea level was rising during interglacial time, the valleys of tributaries which had been eroded in the Tertiary terrain were gradually filled with alluvium, and stream slopes diminished. Consequently, the amount of locally derived bed load carried to the master-stream was greatly

decreased, and heavy minerals transported by the Mississippi became progressively more important, finally forming the dominant suite in the top stratum.

A comparison of heavy minerals in the Tertiary, in terrace deposits, in Mississippi River alluvium, and in the load of the Mississippi River shows a considerable variation in the percentages of each heavy mineral. This is illustrated by the tabulation (table 1) giving percentages compiled from the literature and from studies made for the writer by P. H. Masson, Humble Oil & Refining Company. Some variations noted in the samples studied are related to the grain-size fractions examined, as is indicated by the heavy-mineral content of the two fractions of the suspended load of the Mississippi River (samples N and O, table 1). By destroying certain minerals, weathering also undoubtedly exerts a great influence on mineral percentage, but difference in source is the main factor controlling variations. Because of the large number of source regions drained by tributaries joining the Mississippi in the lower valley, the available mineralogical analyses are insufficient evidence for conclusions concerning characteristic minerals. The analysis of a sample from the Williana formation on Crowley's Ridge near Jonesboro, Arkansas (sample H, table 1), shows a different characteristic suite of heavy minerals than is listed for these gravels elsewhere in the region (Doeglas, 1949; Leighton and Willman, 1949*b*). In this area, zircon, tourmaline, rutile, and staurolite are the more abundant nonopaque heavy minerals, but the sediments are also characterized by an abnormally high percentage of opaque minerals. The use of heavy minerals in establishing the age of lower Mississippi Valley formations through long-distance correlation of a few samples is therefore questionable.

Age and Origin of Lower Mississippi Valley Loess

The many theories advanced to explain the origin of loess have been adequately reviewed by Russell (1944, pp. 6–8) and are well known to those who have given critical thought to the problem. That it is an eolian deposit of Pleistocene age is popularly accepted. This acceptance has been based essentially upon explanation of the physical properties of the loess mass, however, rather than upon faithful description of the conditions under which deposition occurred.

It seems unnecessary to enlarge upon the fact that at the time Russell expressed his views as to the origin of lower Mississippi Valley loess, he had available for his consideration all lines of evidence used in support of the eolian hypothesis. Although the lower-valley loess had all the necessary physical and chemical properties and structure of materials elsewhere considered to be wind deposits, other evidence made the eolian hypothesis untenable. Previously unknown or disregarded facts concerning the composition, distribution, thickness, and topographic expression of loess and its relationship to underlying beds convinced him of the colluvial nature of the deposits.

In reviving the eolian concept for the origin of lower-valley loess, Leighton and Willman introduced evidence based on weathering and mineralogical properties of loess which had not been published at the time Russell presented his theory. Although they recognized the importance of the broad relationships stressed by Russell, they made no attempt either to support or to refute the facts on which he based his theories. They present the theory that the Mississippi alluvial plain was the source area from which loess particles were derived, without considering either

TABLE 1

MINERAL ANALYSES OF LOWER MISSISSIPPI VALLEY LOESS, QUATERNARY AND TERTIARY DEPOSITS, AND MISSISSIPPI RIVER DEPOSITS*

MINERALS	LOCALITIES†														
	A	B	C	D	E	F	G	H	I	J	K	L	M	N	O
Grain Frequency Per Cent Based on Whole Sample before Separation															
Opaque	n.d.‡	.....	.....	.....	.....	.....	.....	3	....	6	4	1	.....	10	.....
Carbonates	42	.....	.....	.....	.....	.....	.....	.....	.....	2	2	2	.....	1	.....
Nonopaque	7	.....	.....	.....	.....	.....	.....	2	....	5	4	3	.....	n.d.	.....
Light	51	.....	.....	.....	.....	.....	.....	95	....	87	90	94	.....	89	.....
Grain Frequency Per Cent Based on Total Heavy Mineral Concentrate															
Opaque	62	18	59	39	.....	.....	60	69	....	28	25	34	.....	38	37
Magnetite	n.d.	.....	.....	.....	A§	A	.....	n.d.	A	n.d.	n.d.	n.d.	.....	n.d.	n.d.
Carbonate	2	.....	.....	.....	.....	.....	.....	.....	.....	17	17	4	.....	2	2
Grain Frequency Per Cent Based on Total Carbonate-free, Nonopaque Heavies															
Epidote	14	29	16	.....	C‖	.....	.....	2	....	6	12	10	22	22	18
Pyroxene group	3	.....	.....	.....	C	C	.....	.....	C	9	12	4	20	16	12
Amphibole (green)	33	39	5	.....	.....	.....	.....	2	C	50#	35	21	33	35	17
Amphibole (brown)	4	.....	.....	.....	.....	.....	.....	.....	.....	.....	6	9	.....	5	6
Apatite	.....	.....	.....	.....	.....	.....	.....	.....	.....	3	4	6	.....	.....	3
Zircon	12	16	30	62	A	A	34	36	A	1	1	10	2	3	12
Titanite	5	2	.....	.....	C	C	.....	5	C	7	7	8	2	6	6
Rutile	3	.....	9	8	C	C	7	13	C	.....	.....	3	<1	5	1
Anatase	.....	1	8	11	C	.....	1	.....	C	.....	.....	.....	.....	.....	2
Staurolite	.....	.....	10	6	C	C	16	14	C	2	.....	.....	<1	2	.....
Topaz	1	.....	.....	.....	.....	.....	.....	.....	.....	2	.....	1	.....	.....	2
Andalusite	3	.....	.....	.....	.....	.....	.....	2	.....	.....	2	.....	.....	2	1
Sillimanite	2	.....	.....	.....	.....	C	4	.....	.....	.....	.....	1	<1	.....	.....
Kyanite	3	1	15	6	C	C	33	3	C	1	1	1	<1	.....	2
Tourmaline	3	.....	2	6	C	A	5	16	C	1	2	6	<1	1	6
Garnet	14	14	5	1	.....	.....	.....	.....	.....	9	6	8	21	3	6
Others**	.....	.....	.....	.....	.....	.....	.....	7	....	9	12	12	.....	.....	6

* Analyses by P. H. Masson, Humble Oil & Refining Company, except where otherwise stated.

† The localities are as follows:

A. Loess, Sec. 9, T. 16 N., R. 4 E., near Vicksburg, Miss.
B. Loess, Hwy. 61, 4 mi. S. of Natchez, Miss. (Doeglas, 1949).
C. Mixed loess (3 ft. below B), 4 mi. S. of Natchez, Miss. (Doeglas, 1949).
D. Upper Tertiary, Hwy. 61, 1 mi. S. of Dolorosa, Miss. (Doeglas, 1949).
E. "Citronelle" gravels, Brookhaven, Miss. (Leighton and Willman, 1949b).
F. Williana formation, Williana, La. (Leighton and Willman, 1949b).
G. Williana sand, Hwy. 61, 0.5 mi. S. of Dolorosa, Miss. (Doeglas, 1949).
H. Williana gravel, Sec. 19, T. 15 N., R. 4 E., N. of Jonesboro, Ark.
I. Bentley formation, Bentley, La. (Leighton and Willman, 1949b).
J. Alluvium, substratum gravels, depth 250 ft., Bayou St. Vincent, La.
K. Alluvium, backswamp clay, depth 17 ft. near Donaldsonville, La.
L. Alluvium, natural levee silt, mile 78, Atchafalaya River, La.
M. Bed load, Mississippi River, Natchez, Miss. (Doeglas, 1949).
N. Suspended load, Mississippi River, Baton Rouge, La. (ret. on 200 sieve).
O. Suspended load, Mississippi River, Baton Rouge, La. (pass. 200 sieve).

‡ Not determined.

§ Abundant.

‖ Common.

Dark-green hornblende distinct from amphiboles.

** Fluorite, monazite, corundum, and undetermined.

the physical conditions which prevailed in the area during glacial stages or the mineralogy of the deposits.

As the present writer feels that the controversy cannot be evaluated properly without an understanding of the main considerations upon which the opposing theories were based, there follows a short discussion of each main topic on which there is a difference of opinion.

DISTRIBUTION OF LOESS

Different usages of the term "loess" have given rise to much confusion as to its distribution; yet this subject is of fundamental importance in considering the question of origin. Russell fully realized that many loesslike materials, such as brown loam and silty loam soils, occur throughout the Mississippi Valley region. For this reason and because he wished to keep any connotation of origin from the definition, he restricted the term "loess" to certain calcareous silts which can be identified on the basis of definite structural, physical, and chemical properties. Leighton and Willman, however, included all silts regardless of the calcium carbonate content.

The calcareous deposits are localized to the outcrop of the Pleistocene terrace formations. They occur on Crowley's Ridge, within the Bluff Hills east of the alluvial valley, and on Sicily Island on the southwestern side of the valley. The noncalcareous "loess," however, is far more common and has a widespread distribution throughout the region.

Regardless of the merit of restricting the term "loess" to calcareous silts, it must be emphasized that if the noncalcareous, loesslike materials mapped throughout the region are considered to be loess, then Leighton and Willman's discussion (1950, pp. 613–614) of the distribution of "loess" in the lower Mississippi Valley is entirely inadequate. Although they use maps and other data prepared by Wascher and others (1948, fig. 1) to show the distribution of the "continuous sheet" of "loess" in the uplands east of the Mississippi, they make no attempt to map or to discuss the significance of the noncalcareous "loess" exposures recorded on the western side of the valley. Wascher and others (1948, p. 396) state that loess occurs at Marksville and Opelousas on the western side of the valley. The location of the towns is shown in figure 2. Loess is recorded in the same general region by Thorp and Smith (1949), who also showed the occurrences on a map exhibited at the El Paso meeting of the Geological Society of America.

It is of interest that noncalcareous "loess" occurs in the western part of the region associated with the 15-mile-wide ancient meander belt of the Mississippi River shown in figure 2. Soils of this meander belt were referred to as loess by Harris (1902, p. 37) and were later mapped as Lintonia and Olivier silt loams by Meyer and Kirk (1915, pp. 1064, 1068), who established their thickness as approximately 2 feet. The noncalcareous "loess" in the ancient meander belt grades down into clay, which borings show to be the upper part of a top stratum of thick clays and interbedded sands and silts with an average thickness of more than 80 feet. Thirty feet of similar materials, called "loess(?)" by Howe and Moresi (1931, p. 124), covers the surface of each of the Five Island salt domes, which rise above the coastal marshlands to elevations of approximately 100 feet.

As there can be little doubt that the noncalcareous "loess" west of the alluvial valley is similar to that mapped on the east, its location should not be disregard-

ed in considering the eolian hypothesis and the theory (Leighton and Willman, 1950, p. 622) that the dominant winds during Pleistocene time were from the southwest. Loess presumably would have been carried into the western uplands by southeast winds.

TOPOGRAPHIC EXPRESSION OF LOESS FORMATIONS

Leighton and Willman accept without reservation the idea (Wascher and others, 1948, p. 394) that the loess is an eastward-thinning wedge, more than 15 feet thick at the Mississippi bluffs and less than 2 feet thick at a distance of 50–75 miles from them. They fail, however, to take into consideration the local distribution and topographic expression of the loess mass which make their thickness figures unrealistic. The location of loess with respect to the topography does not lend support to the idea of eolian deposition, as deposits of this type would be controlled by local wind turbulence, sources of silt, and other factors. Russell (1944, p. 23) points out that calcareous loess covers slopes oriented in all directions and reaches its greatest thickness on valley sides near the foot of the slopes. Obviously, these loess masses must have accumulated without regard to any dominant wind direction. The thickest masses are on the slopes of the most deeply dissected terrain in the region, i.e., the Bluff Hills adjacent to the Mississippi River Valley and Crowley's Ridge, not on the uplands or gentler slopes, where Wascher and others (1948, p. 389) state they made their measurements.

In the photographs used by Leighton and Willman (1950, pl. 1, *B*, pl. 2, *A*), various loess sheets are shown as either horizontal layers or superimposed parallel strata. Nowhere in the text is there any indication that other than conformable relationships exist between the loess "formations." It is strange, in view of the series of alternating erosional and depositional epochs which took place during the Quaternary, that no angular erosional unconformities separate the loess "formations." Leighton and Willman (1950, p. 618) point out that the Peorian loess sheet was laid down on an irregular erosional surface and that it has been deeply eroded since deposition, but they provide no evidence of local erosion separating the Peorian from older loess sheets, nor do they indicate erosional gaps between the older loess sheets.

Some explanation is needed to demonstrate how loess formations could be laid down on steep eroded slopes as parallel layers unaffected by mass movement. Until such an explanation is forthcoming, it should be assumed that the topographic expression of the loess sheets indicates that they all occur as a mantle over the existing topography and hence were not formed during widely separated glacial epochs.

AGE OF UNDERLYING BEDS

In establishing both the age and the origin of loess, it is necessary to take into consideration the age of the beds which the loess material mantles. In the eastern uplands and on Crowley's Ridge, calcareous loess exposures are limited to the terrain of the three highest Pleistocene terraces. The noncalcareous "loess," however, rests upon terrace deposits and the Tertiary uplands east of them. It is recorded by Wascher and others (1948, p. 396) at Delhi, Louisiana, where it mantles the braided stream deposits of late Recent age on Macon Ridge. It is also found on the lowest terrace along the western side of the alluvial valley in southern Louisiana. Thus the noncalcare-

ous "loess" rests on beds of from late Recent to early Tertiary age.

If gravels which the present writer considers to be the basal facies of each terrace formation are still held to be Pliocene, there remains a considerable difference of opinion as to the age of the materials upon which either the noncalcareous or the calcareous loess rests. However, it seems inadvisable to cling to the conclusions of early writers who were without benefit of information now available. The validity of the recent data should have had the fullest attention of Leighton and Willman because their age determinations of loess formations are in large measure dependent upon current interpretations of the age of underlying stratigraphic units. That they recognize this fact is indicated by the following quotation (Leighton and Willman, 1950, p. 614):

> In and near the bluffs the loess deposits generally rest on sand and gravel deposits, commonly called "Lafayette" or "Citronelle" gravel, which the writers consider to be preglacial but which Fisk (1938, 1944) and Russell (1940, 1944*a*) correlate with interglacial stages.

They failed to point out that the age they ascribe to loess formations in some localities is incompatible with the writer's age determinations for underlying beds. In places the loess is given an older age than the beds upon which it rests; i.e., at Memphis, Tennessee, Loveland (Illinoian) loess rests upon the eroded Montgomery (Sangamon) formation; at Natchez, Mississippi, pre-Loveland (pro-Kansan?) loess rests upon the eroded Bentley (Yarmouth) formation.

The literature discloses no factual basis for the preglacial age of sands and gravels underlying the loess. The term "Lafayette" was discarded some years ago when it was found to embrace formations of different ages (see Wilmarth, 1938). The Pliocene age of the Citronelle formation was fully disproved by Roy (1939), who pointed out that the fossil leaves used by Berry (1916) as a basis for his age determinations were obtained at the type locality from clay beds which had been faulted into position. These clays are unconformably overlain by the sands and gravels which are commonly referred to as "Citronelle gravels." Nowhere in the region have definite Pliocene vertebrate or invertebrate faunas been described from the exposed sand and gravel beds. On the other hand, Pleistocene marine invertebrate faunas referred to above (p. 338) date the basal gravels in the deltaic deposits, and Pleistocene vertebrates are found on the outcrop. Elephant and mastodon remains are recorded (McGee, 1891; Hay, 1923) from gravels which form the basal facies of the Bentley formation at Natchez, Mississippi (fig. 3), and which are referred to by Leighton and Willman as preglacial. Elephant remains, together with those of the peccary and the tapir, were found in the terrace gravels in Louisiana by Brown (1938, p. 65). No Pliocene faunas are recognized in the subsurface beds of the deltaic mass near the modern shore line. The first definite upper Miocene faunas occur here at an approximate depth of 8,000 feet, and the only beds which can be referred to as Pliocene are those which occur above this depth and below the basal Pleistocene gravels.

Acceptance of a preglacial age for the gravels involves improbable stratigraphic and topographic relationships. A glance at the cross section (fig. 3), based upon both surface outcrops and data from borings in the latitude of Natchez, Mississippi, shows the distribution of gravels called "Lafayette" or "Citronelle" by Leighton and Willman. The bas-

al gravels of the Williana terrace formation rest on an irregular erosional surface, and their uppermost occurrence in the area is at an elevation of approximately 400 feet above sea level. A lower belt of gravels, the basal facies of the Bentley formation, is present 50 feet above sea level. Similar gravels, forming the lowermost facies of the Recent alluvium, lie more than 200 feet below sea level under the Mississippi alluvial plain. Gravel deposits occur on the western side of the Mississippi Valley at various elevations.

Surely, an age determination of the loess should be accompanied by a reconstruction of the geological development of the area in accordance with known facts. Although they do not reconstruct the physiography and geology of the Natchez area during early Pliocene time, it must be assumed that Leighton and Willman visualize a 100-mile-wide trench cut in Miocene sediments to a depth of 650 feet. This trench would have been filled with a sheet of Pliocene gravels more than 600 feet thick in the latitude of Natchez, Mississippi. In late Pliocene or early Pleistocene, the gravels would have been eroded and replaced by the "Natchez formation" valley-train deposits and later by the fine top-stratum materials of the alluvial valley. The evidence against such interpretations is overwhelming; until the Pliocene age of the gravels can be definitely established, the age determination of the loess must conform with the Pleistocene stratigraphy.

SOURCE OF LOESS

The controversy concerning the age and origin of lower-valley loess makes it necessary to establish accurately a source for the silts. Supporters of the eolian theory believe that the sediments were derived from the Mississippi alluvial plain and were blown to the uplands. According to Russell, however, the loess was derived from weathered backswamp deposits which formed in the top stratum of each terrace formation and accumulated downslope through colluvial action. By this theory, loess is locally derived, and only the local mass movement of the weathered materials is required to bring it to its present position.

Opposition to Russell's field observations is expressed as follows (Leighton and Willman, 1950, p. 620):

> In all these localities the deposits at the top of the section described by Russell as backswamp deposits cannot be differentiated megascopically from the silt in the leached zone, where it directly overlies calcareous loess and is accepted by Russell as weathered loess. No evidence of stratification or textural variation to suggest water deposition was found in any of the deposits described as backswamp deposits by Russell or in scores of similar sections scattered throughout the region. The degree of weathering is not adequate to destroy completely evidence of original bedding, except possibly in the upper 2–4 feet of the deposits, and even here that would probably not be universally true.

For at least one of these localities other published data are available. Vestal (1942, p. 54) described the section of the sediment in the bluff at Natchez, Mississippi, and showed nearly 20 feet of laminated clays beneath the loess, which provide clear-cut evidence of aqueous deposition. These laminated clays are the lower portion of the top stratum of the Bentley terrace formation shown in figure 3. The same aqueous deposit has been mapped and shown diagrammatically by the writer (Fisk, 1944, pl. 4).

OBJECTIONS TO THE MISSISSIPPI VALLEY AS A SOURCE AREA FOR EOLIAN PLEISTOCENE LOESS FORMATIONS

Post–late Wisconsin age of Mississippi alluvial valley.—Appeal to the Mississippi

alluvial valley as the source area for wind-blown silt particles requires that the valley predate the loess "formations." For the Peorian loess to extend to the level of the existing Mississippi Valley surface—the alluvial plain—the present surface must have been in existence since the beginning of Peorian deposition, i.e., the beginning of Leighton's Wisconsin glacial stage. In fact, it must have been in existence as early as the Illinoian glacial stage, as Leighton and Willman (1949*b*, p. 66) record both the Farmdale and the underlying Loveland loess "formations" near floodplain level north of Vicksburg, Mississippi.

Available data appear to demonstrate that neither the Mississippi alluvial valley nor the alluvium which underlies it could have been the source of loess. The valley surface was constructed in late Recent time, and the deposits beneath the plain have all been laid down during and since the waning of the late Wisconsin[2] glacial stage. As well-documented evidence for this in the form of published subsurface information from logs and samples of thousands of borings, faunal studies, and detailed physiographic work (Fisk, 1944, 1947) has been available for some years, it is but summarized here.

Data from the borings prove the Mississippi alluvial plain to be underlain by a thick sequence of alluvial deposits which fills the deeply entrenched pre-Recent valley system. Figure 3 shows the cross-valley relationships of the Mississippi trench and the alluvium in the latitude of Natchez, Mississippi. Considerable evidence supports the late Wisconsin age of the entrenched valley system, which was established from the broad relationships previously discussed (p. 338). Tracing of tributary valleys from the marginal uplands to the master-trench has determined the continuity of the system. Erosion during a single epoch is indicated by the gradual and uninterrupted slope of the valley bottom, which is 200 feet beneath the surface at the northern end of the valley and 400 feet below sea level at the modern shore line. If the slope of the trench is projected seaward to the edge of the continental shelf, it can be seen that the valley system was excavated at a time when sea level was approximately 450 feet lower than at present.

That the entrenched valley system is younger than the youngest Pleistocene coastwise plain is apparent from the fact that it was incised in deltaic facies of the Prairie formation southward from the latitude of Marksville, Louisiana (fig. 2). Invertebrate marine faunas collected from surface exposures by Richards (1939) and from the subsurface by Bridges (1939) show the Pleistocene age of the Prairie formation. The late development and the continuity of the valley system are further indicated by a soil zone which mantles its eroded surface; this zone has been traced directly from the subsurface into the soil mantle on adjacent uplands made up of the Prairie formation and older beds (Fisk, 1944, pls. 4–9; 1947, pls. 60–62). Soil formation resulted from deep weathering caused by the lowering of the water table accompanying stream entrenchment.

The post–late Wisconsin or Recent age of the alluvial filling of the entrenched valley system is indicated by a bed of Recent marine shells lying above the weathered and eroded surface of the Prairie formation beneath the coastal

[2] The term "late Wisconsin" is synonymous with Kay's Wisconsin (fig. 1) and is used here to distinguish the latest glacial stage from Leighton's more inclusive term "Wisconsin."

marshlands. This bed has been traced from 15 feet below sea level in the Lake Pontchartrain area to nearly 100 feet below sea level south of New Orleans (Fisk, 1947, pls. 60–62). Further substantiation of the Recent age is seen in both the gradational characteristics of the alluvium throughout the valley and the physiographic features of the alluvial plain, which together demonstrate that the alluvium was deposited as sea level rose during the waning of the late Wisconsin stage. The steep stream gradients of the glacial stage are reflected in the sands and gravels which make up the coarse basal facies that occur as far south as the modern shoreline. Throughout the valley these coarse deposits grade upward into clean sands and are overlain by a top stratum of relatively fine-grained sands, silts, and clays, which thicken seaward from an average of 40 feet at the north end of the valley to more than 200 feet near the coast.

Low ridges rising above swamplands mark the meander belts of the modern Mississippi River and its older meandering courses. These ridges are continuous throughout the valley and are prominent physiographic features of the alluvial plain. Each of the meander belts has been traced into the deltaic plain, where it splits into minor ridges marking the position of distributary channels. The slopes of both the ancient and the modern meandering river courses are similar and show that changes in position of the Mississippi took place after sea level reached its present stand.

Evidence of earlier stream action can be seen in the northern part of the valley, where alluvial cones bear traces of braided channels of the Mississippi River and its major tributaries. Some of the braided patterns can be traced downvalley into meandering streams of the more gently sloping flood plain. The alluvial cones have been traced southward beneath uppermost floodplain deposits along gradients which indicate that they existed until the sea was within approximately 20 feet of its present level (Fisk, 1944, pl. 18).

The evidence of changing river characteristics from braiding to meandering and the gradual change in character of the alluvium to finer facies indicate that transitional conditions existed in the region until the sea reached its present level. The river systems then established equilibrium between volume, load, and valley slope to permit the poised (Matthes, 1941) condition they have today. Such conditions prove the Recent age of the alluvial mass and indicate that loess could not have been formed earlier than the late Wisconsin, if it originated during glacial epochs and was derived from the Mississippi alluvial-valley region.

Petrography and mineralogy of the alluvium.—The nature, particle size, and degree of sorting of the alluvial sediments make the Mississippi Valley alluvium an improbable source for loess. These features were studied in considerable detail by the writer (Fisk, 1947) in relating them to the environments of deposition created by the river during the rise and final stand in sea level. The positions of natural levees, abandoned channels, backswamp areas, point-bar accretions, and braided streams were established from a study of aerial photographs and maps, and the nature of the associated sediments was revealed by thousands of borings. Many grain-size analyses (Fisk, 1947, pls. 68–70) were made in order to contrast sorting and other characteristics of each environment. Figure 4 shows selected cumulative grain-size frequency curves which were developed in this

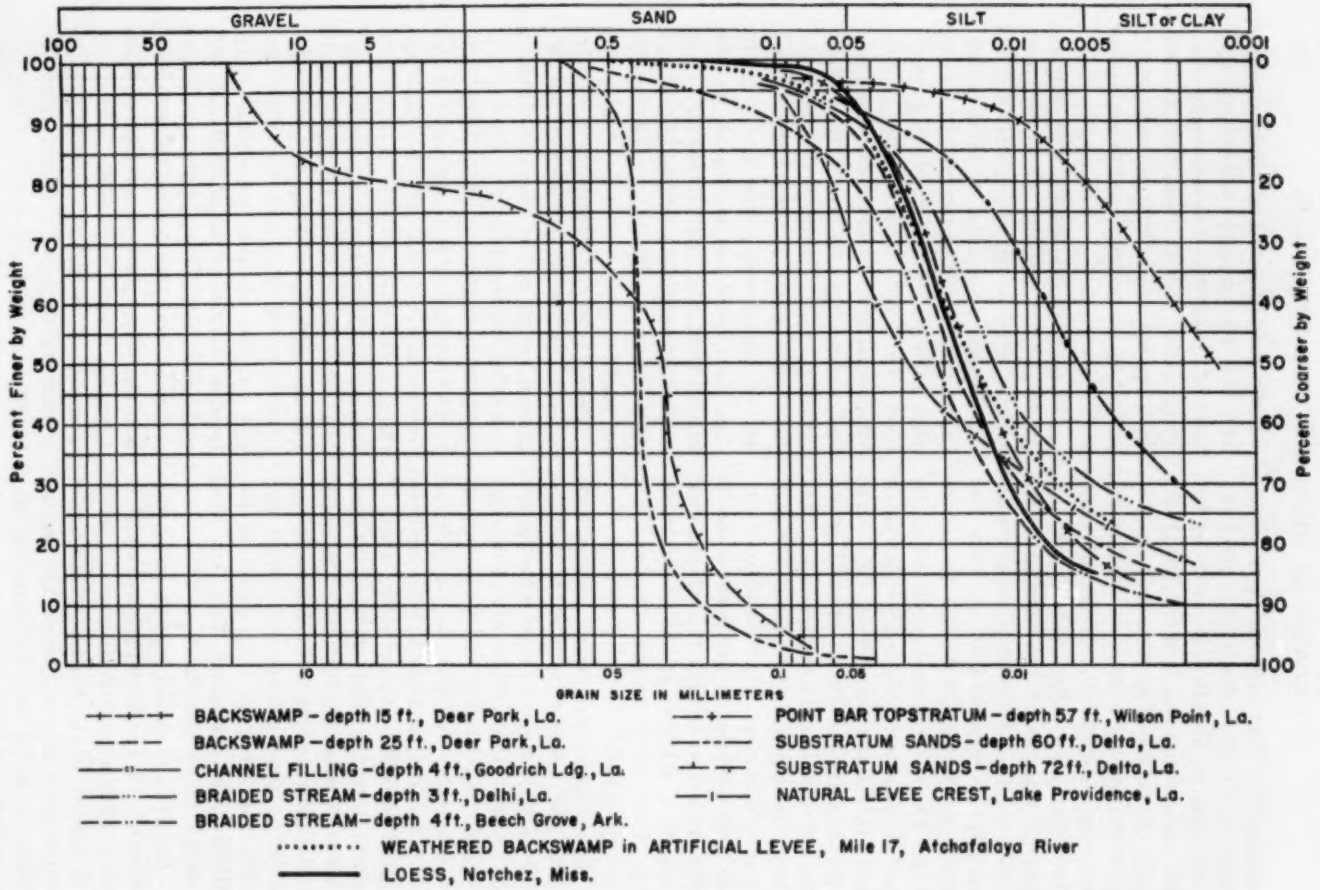
GRAVEL
SAND
SILT
SILT or CLAY
100
50
10
5
1
0.5
0.1
0.05
0.01
0.005
0.001
Percent Finer by Weight
Percent Coarser by Weight
GRAIN SIZE IN MILLIMETERS
BACKSWAMP – depth 15 ft., Deer Park, La.
BACKSWAMP – depth 25 ft., Deer Park, La.
CHANNEL FILLING – depth 4 ft., Goodrich Ldg., La.
BRAIDED STREAM – depth 3 ft., Delhi, La.
BRAIDED STREAM – depth 4 ft., Beech Grove, Ark.
POINT BAR TOPSTRATUM – depth 57 ft., Wilson Point, La.
SUBSTRATUM SANDS – depth 60 ft., Delta, La.
SUBSTRATUM SANDS – depth 72 ft., Delta, La.
NATURAL LEVEE CREST, Lake Providence, La.
WEATHERED BACKSWAMP in ARTIFICIAL LEVEE, Mile 17, Atchafalaya River
LOESS, Natchez, Miss.

FIG. 4.—Cumulative curves of loess and of Recent sediments from depositional environments in the lower Mississippi alluvial valley

work and in subsequent studies of the several environments.

Considerable doubt can be expressed concerning the availability of silt-size particles from the substratum of the alluvium because there are few such grains in the basal gravel-bearing facies. The paucity of silt particles in the substratum is illustrated by the cumulative curves (fig. 4). During the late Wisconsin glacial stage, the gradient of the Mississippi River was steep, and apparently the power of the river to transport suspended silt loads was sufficient to sweep finer particles to the sea. Silts accumulated only late in the history of valley filling; braided stream deposits at the north end of the valley contain abundant silt-size particles, as does the top stratum on point-bar accretions along the modern river (fig. 4). From the standpoint of grain size, the source materials in the alluvial valley would be from the top stratum, which is of late Recent age.

Those who insist that eolian deposits can be distinguished by a remarkable homogeneity in size of particles should examine the sorting of several aqueous environments shown on the cumulative curves of figure 4. It is difficult to distinguish some deposits of braided streams, backswamps, and point-bar accretions from typical loess. In this regard attention is also directed to the similarity between the cumulative curve of loess and that of a sample from a 4-inch weathered zone at the surface of an artificial Atchafalaya River levee constructed of backswamp materials.

Leighton and Willman (1950, p. 621) place considerable emphasis on heavy minerals in differentiating loess formations and use data from Wascher and others (1948, pp. 392–393) and Doeglas (1949) in an attempt to show that the loess could not have been derived from the terrace formations. Analyses were not submitted, however, for the bedded sediments in the top stratum of each terrace, such as those at Natchez, Mississippi, in the upper part of the "Natchez formation." These will be necessary before firm conclusions can be drawn in this respect. The data compiled in table 1 show that the heavy-mineral assemblage in the loess is similar to that in the various environments of the top stratum of the Mississippi River alluvium and to that being carried by the Mississippi River. This mineral suite is distinct from that of the substratum sands and gravels, in that the latter hold dark-green hornblende in great abundance and contain very little epidote and zircon. These marked mineralogical differences, together with the small percentage of silt-size particles, are good indications that loess was not derived during the waning of the late Wisconsin glacial stage, when the substratum sands were being deposited.

Discrete, or primary, grains of calcite and dolomite are believed by Leighton and Willman (1950, p. 618) to make up to 15–30 per cent of unweathered loess. A serious objection to the Mississippi alluvium as a source area for loess is immediately raised because it holds only a very small percentage of carbonate minerals. The results of mineralogical analyses of sediments from the substratum, from natural levees and backswamp deposits in the top stratum, and from the suspended load of the Mississippi River are tabulated as samples J–O in table 1. They show that carbonate grains (including dolomite, siderite, and calcite) average less than 2 per cent of the alluvium by volume. A similar low percentage of these minerals was found in the bed load of the river by R. D. Russell (1937). Obviously, the alluvial sediments could not have provided the necessary high per-

centage of discrete carbonate grains that Leighton and Willman found in the loess.

As there appears to be no other source area for wind-transported carbonate grains, either the carbonates in loess are not primary, or the wind was capable of selectively sorting minerals for transportation. The latter is highly improbable, and it appears advisable to examine further the criteria on which the primary character of the carbonate minerals was established. The writer's study of thin-sections of loess from Vicksburg and Natchez, Mississippi, has not provided proof of primary origin for the majority of calcite particles. The presence of cleavage rhombs and clay inclusions and mass extinction in local areas are petrographic characteristics which suggest the secondary origin of calcite and mechanical disturbance of the loess mass; these support Russell's loessification theory.

Early Recent forests.—In considering the alluvial valley as a source for loess, it is also important to recognize that throughout much of post-Wisconsin time the Mississippi alluvial plain was densely forested; certainly this was so throughout the time that the sea level was rising the last 100 feet. Borings in many parts of the valley south of Vicksburg encounter layers of wood at depths of more than 100 feet, and excavations for cutoffs disclose several superimposed layers of cypress stumps, roots, and logs in the upper 30 feet of the alluvium (Fisk, 1944, fig. 19, *A*). Where the Mississippi and Atchafalaya rivers are actively eroding banks in backswamp deposits, strata containing stumps of trees are exposed more than 20 feet below the surface during low-river stages (Fisk, 1944, fig. 17, *B*). Such a thickly forested plain would have been a poor source area for wind-blown sediments.

STRATIGRAPHIC CONTINUITY OF LOESS

SIGNIFICANCE OF WEATHERING

By including both calcareous and noncalcerous silts, Leighton and Willman are able to designate the surficial material of the lower-valley region as a "continuous sheet" of "loess." This sheet is considered to be of Peorian age because it has essentially the same topographic position, stratigraphic relationships, and general appearance as the Peorian loess of the upper valley (Wascher and others, 1948, p. 390; Leighton and Willman, 1950, p. 617). The only apparent differences are said to be a strong reddish cast of the weathered zone and an increase in depth of leaching from 4 feet in northern Illinois to 15–20 feet in the lower valley.

The idea that leaching has progressed to a given depth throughout the region, irrespective of topography, differences in vegetation, and other variable factors, appears to be the only reason for regarding the surficial material as a definite stratigraphic unit. Leighton and Willman (1950, pp. 618–619) contend that the rate of deposition of wind-blown silts near the bluffs was more rapid than the rate of leaching and that primary carbonate minerals were preserved. According to them, to the east of the bluffs and in other areas where deposits are less than 15 feet thick, deposition was much slower, and the calcareous particles were leached during accumulation. They contend that Peorian loess deposition ceased in Cary time and that subsequent weathering has leached the calcareous loess to a uniform depth of 15 feet.

No older loess "formations" are recognized in the area covered by the noncalcareous "loess," and therefore it must be assumed either that those mapped by

Leighton and Willman in the Bluff Hills were not so extensive as the Peorian sheet or that they were completely destroyed by erosion prior to Peorian deposition. It is also possible that they may be a part of the noncalcareous "loess" classified as Peorian; if so, there can be little stratigraphic significance attached to the Peorian mantle.

In distinguishing the loess "formations" of the Bluff Hills, Leighton and Willman used mineralogical data from Wascher and others (1948, tables 1–7). These data were derived from a study of 27 samples from only four localities in Tennessee and Mississippi, however, and there is little mineralogical evidence upon which to base generalizations concerning the entire region. If minerals are to be considered valid tools for correlating loess, then analyses of the noncalcareous materials east of the Bluff Hills are sorely needed to establish the stratigraphic continuity of the Peorian mantle.

Apparently great care must be used in selecting the area for determining the depth of leaching. Although Leighton and Willman state that everywhere in the lower valley leaching has exceeded 15 feet during and since the last epoch of loess deposition, they cite only one locality near the river bluffs where this is the case (Leighton and Willman, 1950, pl. 2, *A*); other localities pictured (Leighton and Willman, 1950, pls. 2, *B* and *C*) have much thinner leached zones. Furthermore, the observations of the writer do not support the statement that the leached zone thins down the slope as a result of slope erosion (Leighton and Willman, 1950, p. 620). In the many fresh road cuts through the Bluff Hills section, the leached zone is almost everywhere thicker downslope than at the crest of the cuts (for substantiation see photograph by Leighton and Willman, 1950, pl. 2, *C*). Indications of stream erosion are difficult to find on the densely forested, loess-covered slopes, which are underlain by thick fertile soils. Assertions that the original Peorian loess sheet is now deeply eroded would be greatly strengthened by evidence.

The writer believes that the use of weathering and related phenomena as a means of interregional classification should be seriously questioned until the relative rates of weathering in such dissimilar climatic regions as the upper and lower valleys have been accurately determined both for the present and for the Pleistocene. In the upper valley, where each loess formation is partly covered by drift sheets of known age, the depth of weathering which took place before burial of the loess can be measured. In the nonglaciated lower valley, however, weathering progressed without interruption. Until the stratigraphic position of weathered materials can be accurately determined, it is impossible to establish a factual basis for distinguishing depth of weathering for significant time intervals.

SIGNIFICANCE OF COLLUVIATION

Leighton and Willman vigorously oppose Russell's loessification theory, but they do not refer to the colluvial phenomena on which it is largely based, although they visited many of the localities from which Russell made his observations and no doubt saw the features he described. The field evidence shows that colluviation has been in progress throughout the period of loess formation; the orientation of gravel particles in the basal part of the section provides proof of early mass movement, and the multitude of minute terraces which everywhere

characterize loess slopes show that the process is continuing. As these features are not mentioned, it may logically be asked whether Leighton and Willman gave any consideration to the effect which colluvial processes might exert on the local rate of weathering and whether they investigated the possibility that the superimposed loess "formations," which they describe as representing distinct stages of eolian deposition, might actually represent separate epochs of colluviation.

The effects of colluviation are indicated in the heavy-mineral content of loess, as shown in published analyses (Doeglas, 1949, table 2) which are listed as samples B and C in table 1. Sample C from a lower layer in a loess exposure on Highway 61, 4 miles south of Natchez, Mississippi, is called "mixed loess" by Doeglas. In addition to the heavy-mineral suite of hornblende, epidote, and garnet which is typical of loess elsewhere in the region (Wascher and others, 1948, tables 1–7) and of the local loess 3 feet higher in the section (sample B), the "mixed loess" samples contain a high percentage of kyanite, staurolite, and zircon. This latter suite of heavy minerals characterizes the Williana gravels, which the loess locally overlies. The lower layer also contains gravel particles which give additional clues as to the effect of colluviation.

Mixing of loess with other materials through colluvial action is known. In one case cited by Russell (1944, p. 14) lignite was traced downslope from its outcrop into fragments present in loess. Mixing is also indicated by the presence of fresh-water mollusks in loess near Vicksburg, Mississippi. Mr. Charles R. Kolb, geologist, Waterways Experiment Station, has collected specimens of the snail *Lymnaea* and the clam *Sphaerium*, widely distributed fresh-water genera, from an exposure in the southeast corner, Sec. 9, T. 16 N., R. 4 E. These fossils were present in calcareous loess approximately 30 feet above its contact with the basal sands and gravels of the Bentley formation. The presence of only separate valves of the clam, together with their haphazard distribution, proves that the shells have been shifted from their original positions. Similar fresh-water mollusks have been described by Richards (1938, pp. 28–29) from Pleistocene silty sands underlying loesslike material along Little Bayou Sara, East Feliciana Parish, Louisiana, and from several other localities in southern Louisiana and Mississippi.

If the effects of colluviation are not recognized, then the existence of fresh-water mollusks and the mixed suites of heavy minerals in some loess present intriguing problems. If such effects are taken into consideration, they present formidable obstacles to those wishing to establish the stratigraphic continuity of loess "formations" in the lower valley.

OTHER CONSIDERATIONS

Geologists advocating the eolian theory of formation for lower Mississippi Valley loess should be prepared to provide a logical reason for the absence in the region of sand-dune deposits of comparable age. The present river is depositing huge sand bars, and, as has been pointed out, the great bulk of the substratum is sand. In the course of many river trips the writer has observed that, although winds commonly shift sand on bars and pile it up in small dunes, bedded silts, which also occur on the bars, are not subject to wind transportation. Even when dry conditions prevail, the moisture contained in silt is apparently sufficient to make it extremely resistant to wind erosion. Pleistocene winds strong enough to abstract

and blow silt-size dust particles for such distances as 75 miles from the river should have been able to shift the sand into dunes and pile it against the bluffs. The absence of sand-dune features leads to serious doubt as to the local effectiveness of wind as a carrying agent; as a matter of fact, the existence of sand dunes should be taken into account wherever loess is thought to have been derived from glacial outwash, which also contains large quantities of sand-size particles.

Leighton and Willman (1950, p. 605) state that the fauna of loess indicates a forested region. If wind blew dust into forested uplands, where are the trees that it buried? Surely, some river bluff or road cut in the lower-valley region would expose a preserved log. Inasmuch as no stumps or logs have been revealed in the deposits, trees must not have been buried by loess. It seems more logical to entertain Russell's hypothesis of loessification, which does not demand burial of surface features; colluviation phenomena involve soil creep and other slow downslope movements which disturb surface features slightly but are more pronounced in the subsurface.

Conclusions

It is doubtful that the eolian hypothesis as applied to the lower Mississippi Valley loess has been strengthened by the studies of Leighton and Willman or by those of Wascher, Humbert, and Cady. Its proof there requires far more detailed studies than have been presented and the introduction of a completely new and more logical line of reasoning. An acceptable explanation must take into account the geological setting and broad regional relationships and be consistent with the physical and chemical properties of loess. It should benefit from the wealth of data on the Quaternary geology of the region which have been obtained during the last twenty years.

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HISTORIC AND PREHISTORIC FLUCTUATIONS OF ALPINE GLACIERS IN THE MOUNT GARIBALDI MAP-AREA, SOUTHWESTERN BRITISH COLUMBIA[1]

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ABSTRACT

Historical, botanical, and geological studies in a part of the Coast Mountains of British Columbia show that all the existing glaciers of that area are retreating from climaxes attained in the early part of the eighteenth century and in the middle of the nineteenth century. A rapid rate of shrinkage, amounting to a loss in depth of as much as 12½ feet of ice annually, has prevailed, however, only since the second and third decades of this century and has coincided approximately with a relatively warm period indicated by meteorological records. In three of four glaciers studied, the maximum advance of the last three centuries exceeded any other since the latter part of the Pleistocene epoch, and one of these three glaciers has been as extensive as it was in 1947 for only about four centuries since the deglaciation of its basin by the Cordilleran ice sheet in late Wisconsin time. For a fourth glacier, however, the advances of the past few centuries failed by far to reach the limits of a much earlier, probably late Pleistocene, advance. Reduction in its potential accumulation area as a result of unusually rapid lowering of its cirque head wall by erosion may account for the anomalous behavior of this glacier. Many low-level cirques, formed by early or pre-Wisconsin alpine glaciation, have not been reoccupied by local bodies of ice during or since the disappearance of the Cordilleran ice sheet.

INTRODUCTION

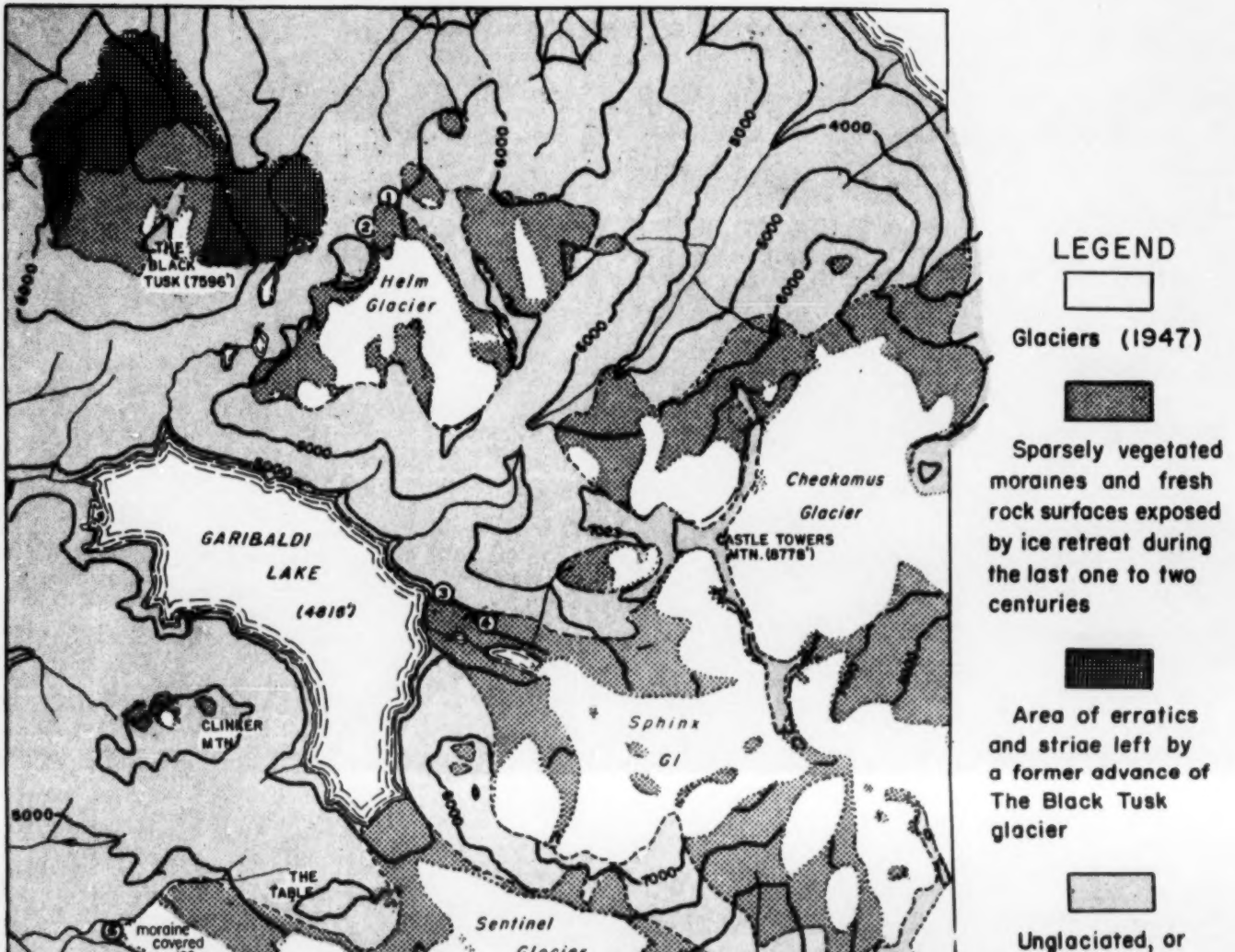
Data accumulated in the course of geological studies in the Mount Garibaldi map-area of southwestern British Columbia indicate the extent of alpine glaciers at several stages in Pleistocene and post-Pleistocene time. Historical records, measurements, and photographs can be used to assess changes in the glaciers for only the brief interval since the first explorations of the alpine part of the map-area in 1906–1912. Botanical data provide a fairly satisfactory dating of two climaxes of the mountain glaciers within the last few centuries, and geological data give clues to the extent of the glaciers at still earlier periods.

DESCRIPTION OF THE AREA

The Mount Garibaldi map-area (fig. 1) lies in the southern Coast Mountains and extends from 25 to 50 miles north of the city of Vancouver. The area is rugged, ranging from sea level to an altitude of 8,787 feet.

During the Pleistocene epoch the map-area was covered by the Cordilleran ice sheet, which left erratics and striated surfaces up to altitudes of 7,200 feet in the northern part of the map-area and to somewhat lower altitudes in the southern part. Evidence of only one period of regional glaciation has been found within the Coast Mountains, although two distinct till sheets separated by fossiliferous interglacial beds are known at Vancouver (Johnston, 1923). Judging from the freshness of the till, erratics, and glaciated surfaces in the map-area, it is the Wisconsin glaciation that is recorded there, traces of earlier glaciations being presumably either buried or eroded away. During the climactic and postclimactic stages of the ice sheet the general direction of ice movement was a few degrees west of south across the map-area. The distribution of erratics and trends of striae, however, indicate local centers of ice dispersal in the vicinity of the higher massifs, such as Castle Towers Mountain

[1] Manuscript received June 26, 1950.


THE BLACK TUSK (7596')
Helm Glacier
Cheakamus Glacier
CASTLE TOWERS MTN. (8775')
GARIBALDI LAKE (4618')
CLINKER MTN.
Sphinx Gl
THE TABLE
moraine covered
Sentinel Glacier
4000
5000
6000
7000
8000
LEGEND
Glaciers (1947)
Sparsely vegetated moraines and fresh rock surfaces exposed by ice retreat during the last one to two centuries
Area of erratics and striae left by a former advance of The Black Tusk glacier
Unglaciated, or

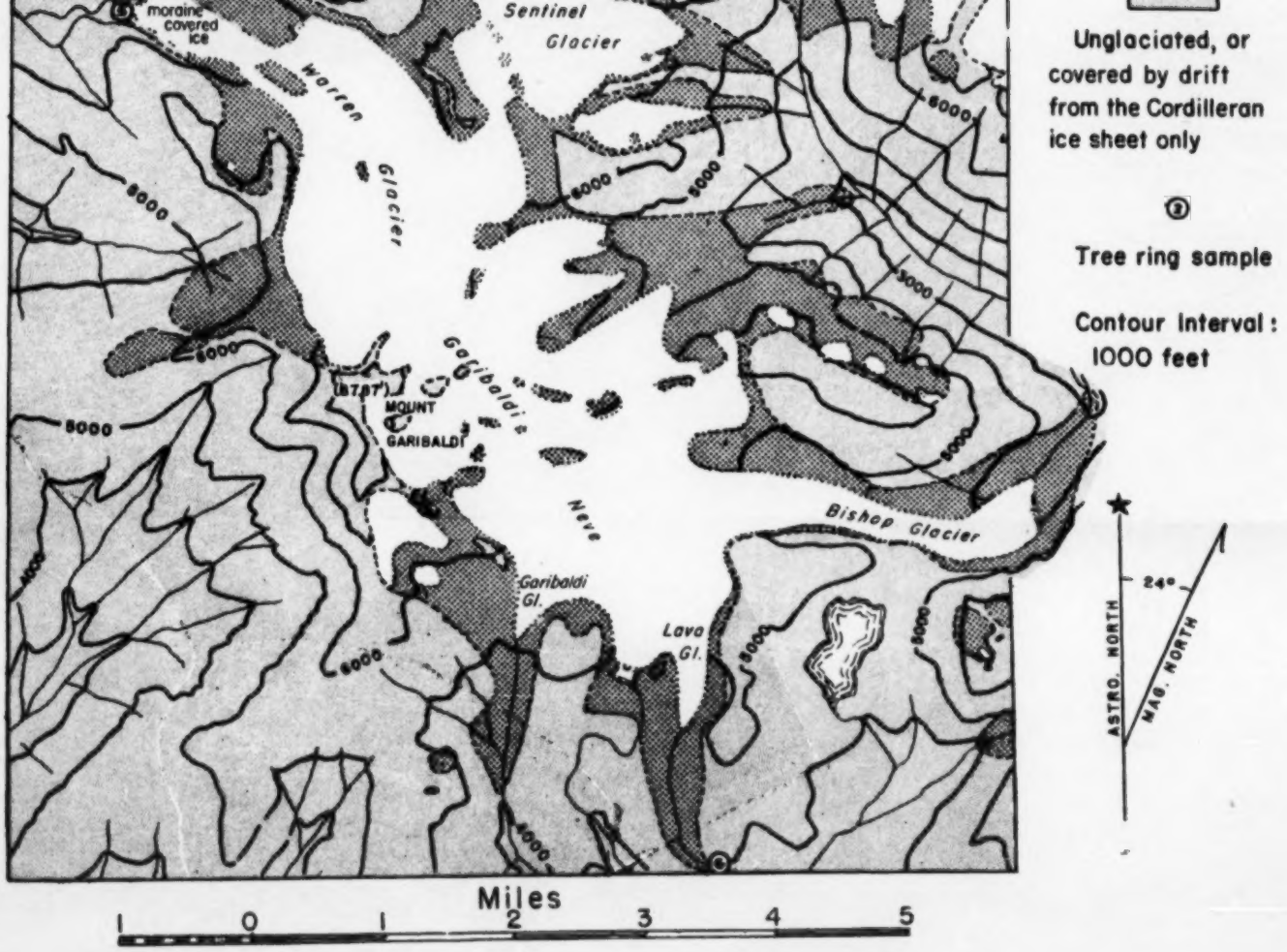
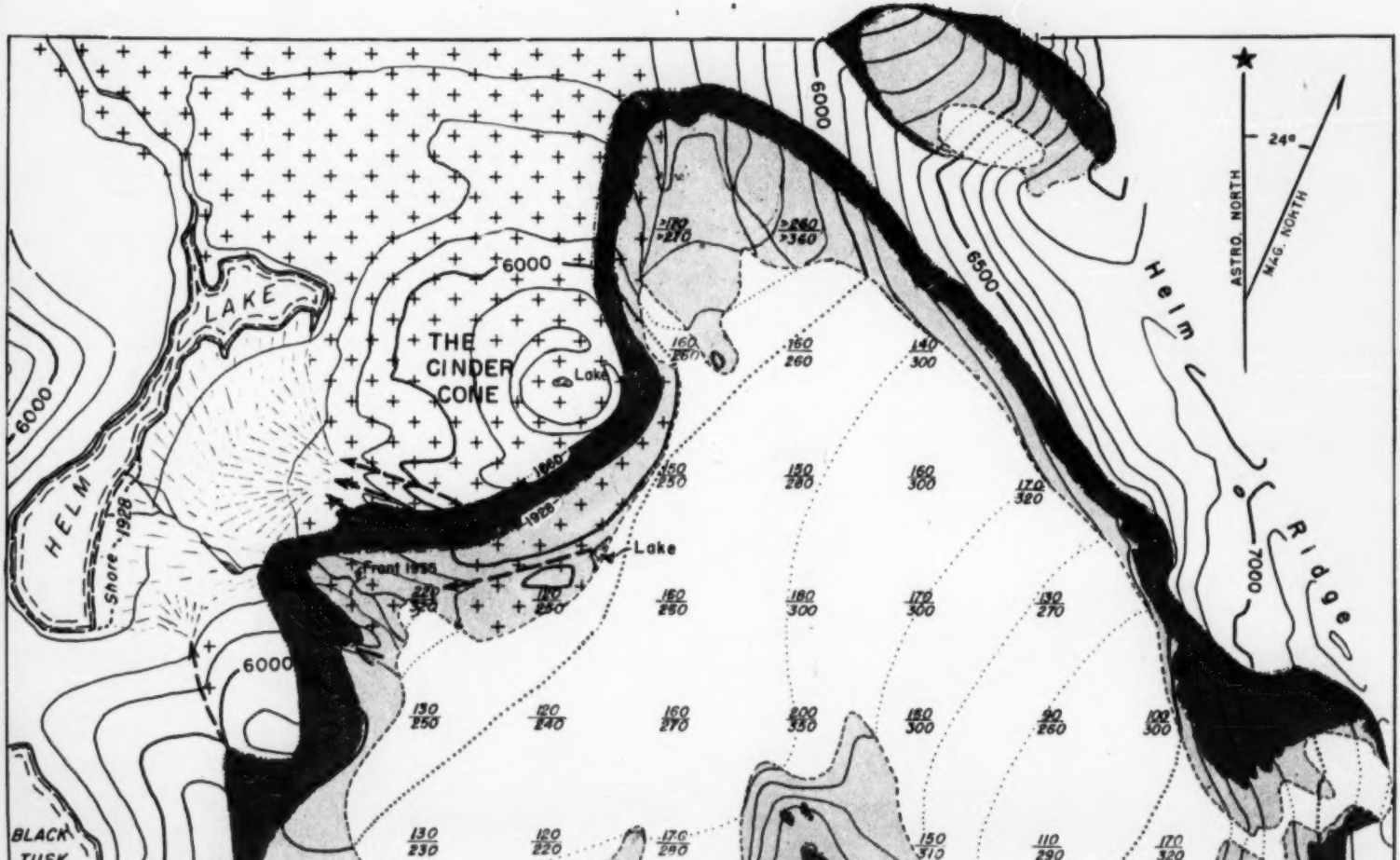
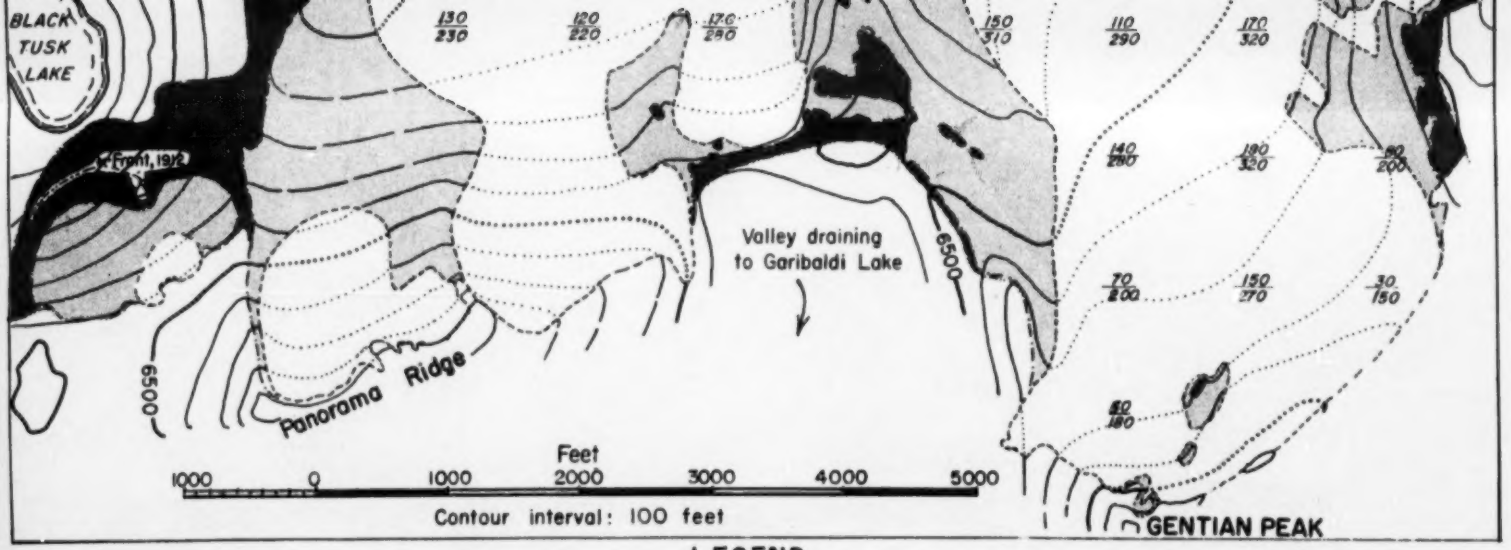


FIG. 2.—Map of the northeastern part of the Mount Garibaldi map-area, showing changes in alpine glaciers


6000
6500
7000
Helm Ridge
LAKE
HELM
THE CINDER CONE
Lake
Lake
Shore 1928
Front 1935
1928
1935
ASTRO. NORTH
MAG. NORTH
24°
BLACK TUSK



LEGEND

Glacier and firn in 1947

Area exposed by ice recession from time of recent climax (ca. 1860) to 1928

Area exposed by ice recession from 1928 to 1947

Area unglaciated since disappearance of Cordilleran ice

Abandoned meltwater channels

Alluvial fan

Ash and lapilli, lava

150/300 Drop of ice surface in feet 1928–1947
Drop of ice surface in feet climax–1947
Values considered correct to within 25 ft.

FIG. 3.—Helm Glacier

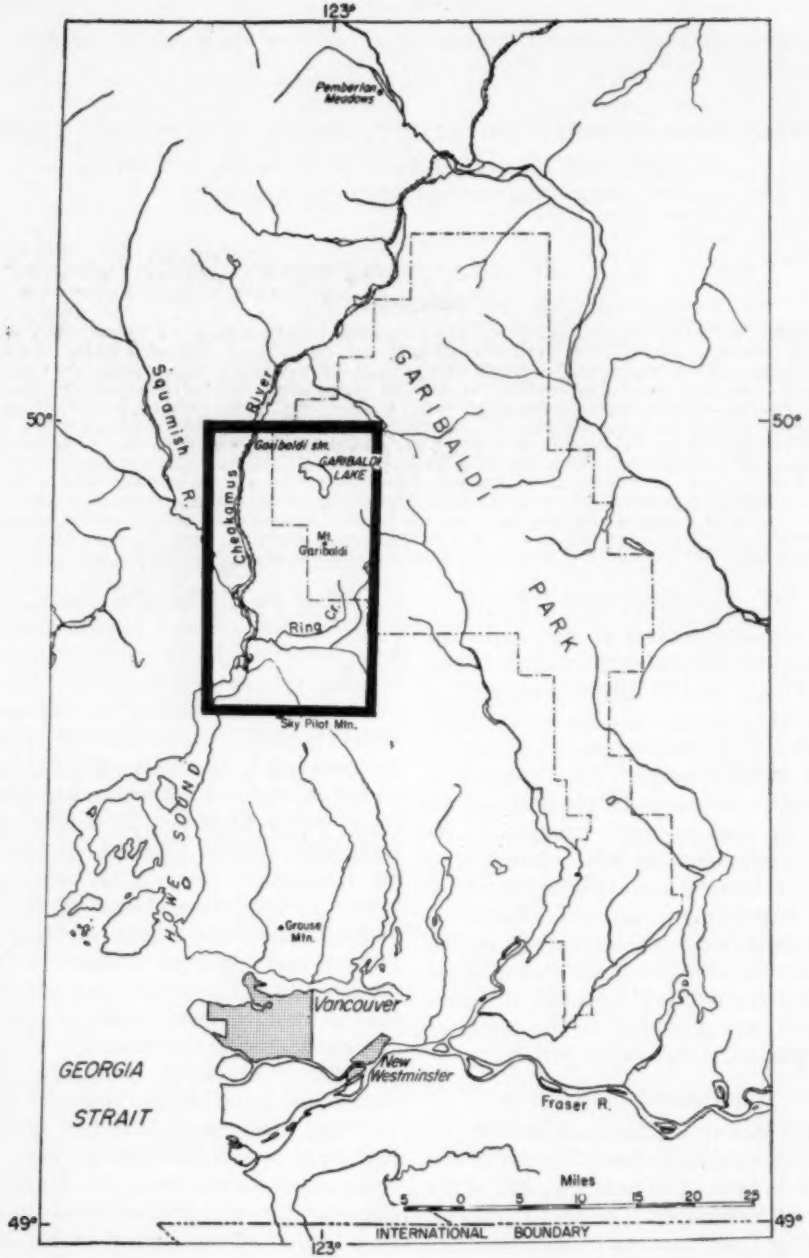


FIG. 1.—Index map, showing location of the Mount Garibaldi map-area

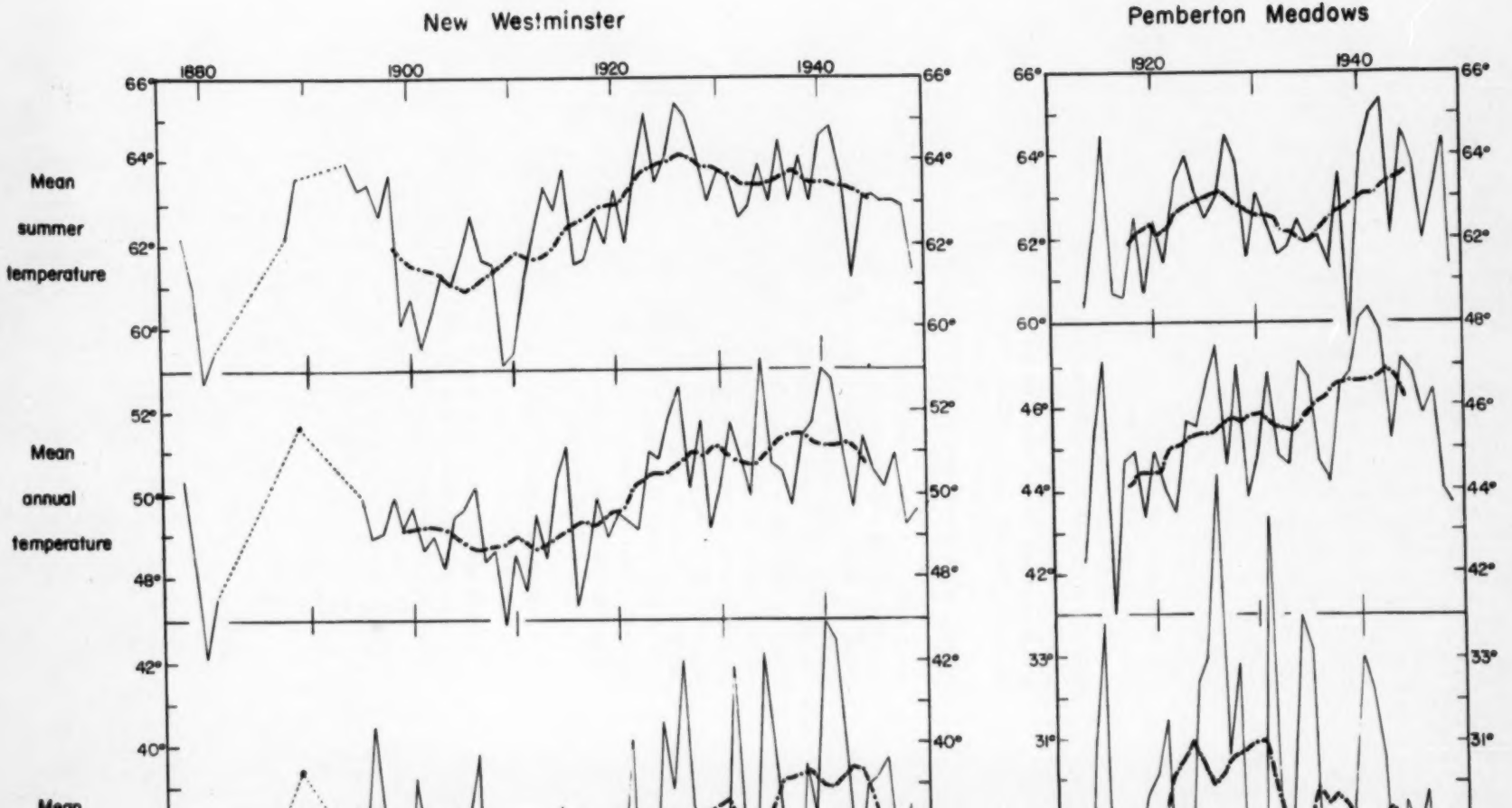
(altitude 8,778 feet), from which ice moved westerly and perhaps even northerly for short distances. Such massifs can be regarded as nunataks projecting above the general level of the ice sheet and supporting local glaciers which were confluent with the ice sheet itself. The drift transported and deposited by the Cordilleran ice sheet is of mixed composition, reflecting the varied lithology of the region, but boulders of quartz diorite and of greenstone—the two predominant and resistant rocks of the Coast Mountains—are conspicuous.

The time at which the last of the Cordilleran ice disappeared from the Mount Garibaldi map-area is not known precisely. Isostatic emergence of the fiord coast of southern British Columbia was, however, still incomplete when the ice was disappearing from the head of Howe Sound, 13 miles southwest of Mount Garibaldi; for in this locality clays postdating glacial deposits and containing casts of marine pelecypods occur up to an altitude of 100 feet. Johnston (1921), from his studies of the growth of the Fraser River delta, concluded that approximately 8,000 years had elapsed since emergence was complete and the construction of the 250 square miles of delta below New Westminster began. His estimate, however, does not include the time required for Fraser River to fill with sediment several large inland basins above New Westminster, totaling 275 square miles. Emergence, therefore, was probably completed more than 10,000 years ago, and deglaciation in the Mount Garibaldi map-area at a still earlier date.

Although most of the Mount Garibaldi map-area is underlain by granitic and metamorphic rocks, Pleistocene volcanic rocks are locally important, and some of the mountains, notably Mount Garibaldi (fig. 2), The Table, Clinker Mountain, The Black Tusk, and The Cinder Cone (fig. 3), are built largely of these extrusives. As far as can be ascertained, volcanism ceased about the time of disappearance of the last ice sheet. One of the latest eruptions, centering at Clinker Mountain, gave rise to two large lava flows, which at higher levels present normal surface gradients and flow structures but which at their lower ends were ponded to depths of at least 800 feet, apparently against the waning Cordilleran ice sheet when it still covered all slopes below the 4,000-foot level in this part of the map-area.[2] Another eruption from a center 2 miles southeast of the summit of Mount Garibaldi gave rise to the "Ring Creek lava flow," 11 miles in length. Although apparently no part of the latter flow came in contact with relics of the ice sheet, indirect evidence[2] indicates that ice was still lingering in, or had very recently disappeared from, Squamish River Valley, near the terminus of the flow, when the lava came to rest.

The climate of the area is today sufficiently cool and moist to support glaciers at higher altitudes. Precise data on the climate are scanty, the only official meteorological records from within the map-area having been made at the settlement of Garibaldi (formerly Daisy Lake) (altitude 1,100 feet) from 1922 to the present and at Garibaldi Lake (altitude 4,800 feet) for a period of eleven months in 1933. Mean annual precipitation at Garibaldi up to 1947 was 59.95 inches, more than half of this falling in four winter months. Precipitation at Garibaldi Lake for the eleven-month period was almost one and two-thirds that at Garibaldi; hence an average of

[2] Unpublished Ph.D. thesis of W. H. Mathews, "Geology of the Mount Garibaldi Map-Area," at the University of California library, Berkeley, California. Report for publication in preparation.


New Westminster
Pemberton Meadows
1880
1900
1920
1940
Mean summer temperature
Mean annual temperature
66°
64°
62°
60°
52°
50°
48°
42°
40°
46°
44°
33°
31°

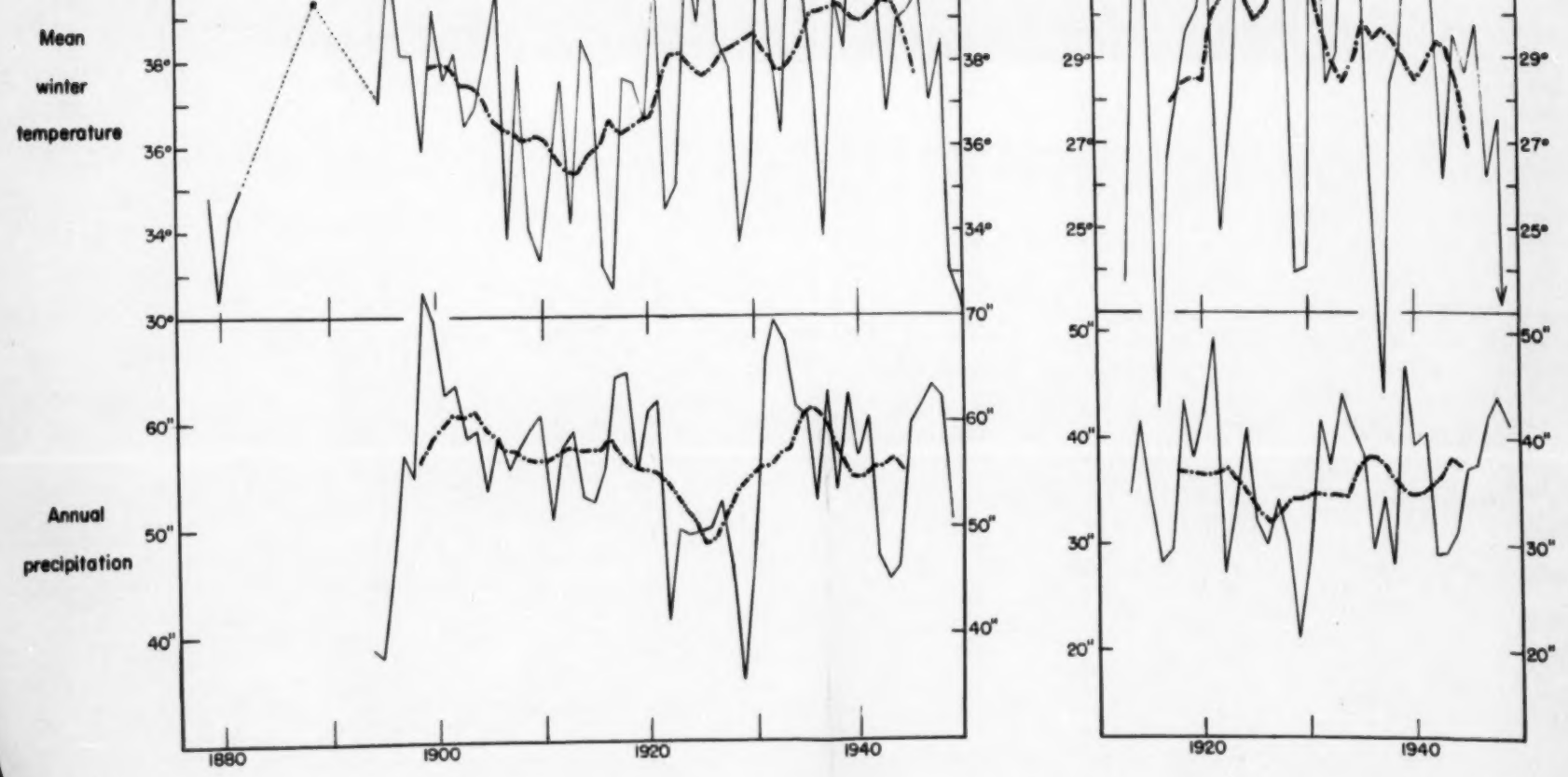


FIG. 4.—Annual temperature, annual precipitation, and moving ten-year averages (*heavy broken lines*) at New Westminster and Pemberton Meadows, British Columbia

95–100 inches annually might be expected at the former locality. Streamflow observations on Cheakamus River at Garibaldi and on the stream draining Garibaldi Lake indicate a runoff of slightly more than 80 inches a year from their drainage basins (Dominion Water and Power Bureau data). Of this runoff, an amount equivalent to about 5 inches in the former stream and about 10 inches in the latter is contributed annually by surplus melting of ice from glaciers within their basins. Loss of precipitated moisture through evaporation may be between 15 and 30 inches a year, judging from studies in comparable areas (Lee, 1942, p. 321). The average annual precipitation within the drainage basis is therefore estimated to be about 100 inches a year, ranging from as low as 60 inches a year in some parts to an undetermined amount above 100 inches in others. The mean annual temperature at sea level at this latitude in British Columbia is 50° F. Judging from the incomplete records at Garibaldi Lake, the mean temperature for 1933 was about 34° in a year which elsewhere was 1° below the average for the 1930's (fig. 4). The mean summer temperature at Garibaldi Lake in 1933 was 48° in a period which was 0°.35 below the average for the same decade. The temperature at The Cinder Cone (altitude 6,000 feet) has been sufficiently cold for the preservation of permafrost to within 2½ feet of the surface on a sheltered exposure. This permafrost extends to a depth of at least 40 feet below the surface in the western part of this cinder deposit; hence cold temperatures have prevailed here for a long period, perhaps since the last ice age. Because of cool winter temperatures and high winter precipitation, the snowfall is heavy. Snow accumulated to a maximum depth of 25 feet in the spring of 1946 at a point 4 miles south of Mount Garibaldi and 4,800 feet above sea level. Snowfall north of Garibaldi Lake at similar altitudes is believed to be somewhat less than on the slopes south of Mount Garibaldi. Near Garibaldi Lake the firn line is between 6,500 and 7,000 feet above sea level, and ridges as low as 6,700 feet may support glacierets on their northern slopes. Farther south the firn line is lower, and Sky Pilot Mountain (altitude 6,645 feet), 15 miles south of Mount Garibaldi, bears on its northern face a small glacier which extends down to about the 5,000-foot level.

More than two dozen alpine glaciers and glacierets, having an aggregate area of slightly more than 22 square miles, existed in 1947 within the area encompassed by the present study. With the exception of the one glacier on Sky Pilot Mountain, all were clustered in the northeastern part of the Mount Garibaldi map-area (fig. 2), where most of the higher peaks are to be found. The largest individual glacier, the Cheakamus, had an area of about 3¼ square miles, but Garibaldi Névé and its distributaries, together with Sentinel Glacier, formed a continuous body of ice about 11½ square miles in area.

RECENT RECESSION OF THE ALPINE GLACIERS

A recent recession of all the glaciers of the map-area is clearly demonstrated not only by historic records and photographs but also by the zone of barren to sparsely vegetated moraines and fresh rock surfaces surrounding them. The recent limit from which this recession has taken place is clearly marked in front of nearly all the glaciers by a sharp boundary, the so-called "trimline," between mature timber and herbaceous or shrubby vegetation. A conspicuous "high ice mark,"

separating nearly barren rock or moraine below from timbered, heath-covered, or lichen-coated surfaces above, is readily apparent along the walls of many valleys from a few tens to several hundred feet above the level of the glacier.

The time at which recession began and the initial rate of retreat can be ascertained only from botanical and geological records because the glaciers had withdrawn a few tens to a few hundreds of feet from their limits when the first mountaineers visited the area. On Helm Glacier the time of the climax has been inferred as about 1860 (p. 366), and, judging from the similarities in the stage of re-vegetation of moraines, the maximum extent of most of the other glaciers was reached about the same time. In spite of the lack of a precise date for this ice limit, it serves as a useful reference line in the measurement of subsequent retreat.

The rate of retreat of glacier termini, the only measure of recession commonly available in the records, is in most instances markedly influenced by irregularities of the glacier floor. Both the west tongue of Helm Glacier and the northern half of Warren Glacier, however, end on relatively flat floodplains, and in these two examples the lineal retreat is a fair measure of changes in the glacier regime. The same is true for the Sphinx Glacier from the time its front ceased to be in contact with the waters of Garibaldi Lake (an indefinite time prior to 1912) until it withdrew to the inner end of its outwash plain in 1946. The lineal retreat of these three glaciers in recent years, computed from old photographs with supplementary data provided by W. Taylor and the Dominion Water and Power Bureau, is given in table 1.

A fall in the level of the ice surface is clearly indicated in photographs taken from the same point at different times over a period of several decades (pls. 1; 2, *A*). Such photographs indicate, for example, that the surface of Garibaldi Névé near the 5,500-foot level east of Mount Garibaldi has been lowered by an averge of 165 ± 25 feet in the 36-year period 1911–1947, that is, at a mean rate of 4½ feet annually. The southern distributaries of this névé have become as much as 300 feet thinner in the same period, that is, at a mean annual rate of about 8 feet. The northern part of Warren Glacier near its lower end has been decreasing in thickness at annual rates of up to 12½ feet a year (table 1).

The most complete data on recent recession and thinning of the ice are based on photographs of Helm Glacier taken in 1928 (pl. 2, *B*) by A. J. Campbell, of the British Columbia Department of Lands and Forests, and again in 1947 by myself from the same camera stations. From these photographs it has been possible to prepare detailed maps showing the position and altitude of the ice margin at three stages: (1) at the time of the climax, about 1860, (2) in 1928, and (3) in 1947. From these maps it has been possible to draw approximate contours on the ice surface at each stage and to estimate thinning of the ice between successive stages (fig. 3). A comparison of the three maps reveals that the area of this glacier declined from a maximum of 56,400,000 square feet about 1860 to 48,800,000 in 1928 and to 34,300,000 square feet in 1947. Three small masses of ice had become partly or completely isolated from the main glacier during this shrinkage, and one of these, at the western end of Panorama Ridge, had almost completely disappeared. The volume decrease of the glacier prior to 1928 is estimated to have been at least 6,000,000,000 cubic feet, representing an average loss

TABLE 1

GLACIER RETREAT IN THE MOUNT GARIBALDI MAP-AREA

GLACIER	PERIOD	RECESSION		AUTHORITY*
		Total (Feet)	Mean Annual (Feet)	
Helm Glacier, west tongue, retreat of ice front	1860±–1912	150± 50	3±	WHM
	1912 –1928	200± 50	12½	WHM
	1928 –1935	400± 50	57	WHM
	1935 –1936	46	46	WT
	1936 –1937	21	21	WT
	1937 –1945	293	37	DWP
	1945 –1946	21± 11	26	DWP
	1946 –1947	52± 29		DWP
	1947 –1948	19± 8		DWP
	1948 –1949	11		DWP
Helm Glacier, east edge at base of Helm Ridge, lowering of ice surface	1860±–1913	130± 15	2.5± .3	WHM
	1913 –1928	40± 10	2.7± .6	WHM
	1928 –1947	150± 10	8 ± .5	WHM
Warren Glacier, retreat of ice front	Climax–1900±	1,150±100	.....	WHM
	1900±–1906	100± 20†	15	WHM
	1906 –1912	240± 20†	40	WHM
	1912 –1922	600± 50†	60	WHM
	1922 –1928	640± 50	110	WHM
	1928 –1935	700±100	100	WHM
	1935 –1936	80	80	WT
	1936 –1938	208	104	WT
	1938 –1947	1,102	122	WT
Warren Glacier, south of the east end of The Table, lowering of ice surface	Climax–1922	165± 15	.....	WHM
	1922 –1928	60± 15	10 ±1.5	WHM
	1928 –1946	225± 20	12.5±1	WHM
Sphinx Glacier, retreat of ice front	Climax– ?	Retreat of a few hundred feet from deep water to the shore of Garibaldi Lake		
	? –1912	300± 50	?	WHM
	1912 –1928	1,600±100	86	WHM
	1928 –1935	1,200±100	170	WHM
	1935 –1936	220± 40	140	WT
	1936 –1937	60± 20		WT
	1937 –1947	1,440±100	144	WHM

* Authorities: W. H. Mathews (WHM); W. Taylor (WT); Dominion Water and Power Bureau (DWP).

† Measurements based on a series of fairly regularly spaced and uniformly sized moraines, believed to represent annual accumulations.

PLATE 1

A, Warren Glacier from Clinker Mountain, 1912. Photo courtesy of W. Taylor.

B, Warren Glacier from Clinker Mountain, 1928. Photo by Topographic Division, British Columbia Department of Lands and Forests.

A

B

Warren Glacier, 1912 and 1928

A

B

Warren Glacier, 1945, and Helm Glacier, 1928

in thickness of between 100 and 130 feet over the extent of the glacier. The volume decrease of the shrunken glacier between 1928 and 1947 is similarly estimated to have been about 5,000,000,000 cubic feet, representing an average loss in thickness of between 120 and 150 feet, or a mean annual loss of between 6 and 7 feet over the entire area of the glacier. As might be expected, the loss in thickness has been greater in that part of the glacier below the firn line than in the névé areas, but even in the latter part it has shrunk materially.

Measurements of year-by-year accumulation surpluses or deficits in various parts of the glacier have been provided by Taylor (personal communication) and are given in table 2, with comparative data on snow depths on Grouse Mountain, near Vancouver (fig. 1), and mean summer temperatures at New Westminster and Pemberton Meadows. The observations demonstrate the wide variations in the rate of accumulation of firn in those parts of the glacier above snow line and the relatively small changes in the rate of ablation in the lower part of the glacier. Annual surpluses or deficits in the Helm firn field show a correlation with the maximum depth of snow accumulated the previous winter on Grouse Mountain. Had measurements of maximum snow depth been made in the immediate vicinity of Helm Glacier, a still closer correlation might have been obtained. Variations in the amount of snow drifting into this and other parts of the Helm firn field from adjacent slopes within and beyond the drainage basin

TABLE 2

WASTING AND ACCUMULATION ON HELM GLACIER AND CORRESPONDING SPRING SNOW DEPTHS AND SUMMER TEMPERATURES

Year	Drop in Level of Ice Margin at Southeastern Base of the Cinder Cone* (Inches)	Surplus on Margin of Helm Névé, about ½ Mi. Northeast of Gentian Peak (Feet)	Maximum Snow Depth at Grouse Mt. (Feet)	Mean Summer Temperature (° F.)	
				New Westminster	Pemberton Meadows
1938	73	+6	15	64.1	63.6
1939	69±	+7	18½	63.0	59.7
1940	74	−7½	7	64.6	64.0
1941	68	−6	7½	64.8	65.0
1942	72	−4	9½	63.7	65.4
1943	67±	+5	13	61.2	62.1
1944	73	No data	No data	63.1	64.7

* The figures quoted indicate the drop in the level of a point constantly shifting with the retreat of the ice margin. They are about 20 per cent lower than the figure that would be obtained in this locality for the drop at a fixed point on the ice surface.

PLATE 2

A, Warren Glacier from Clinker Mountain, 1945. Photo by Roy Howard.

B, Helm Glacier, the Cinder Cone, and Helm Lake delta from the summit of The Black Tusk, 1928. The advance of the delta front has been so rapid that the narrow strip of Helm Lake visible from The Black Tusk in 1928 (*arrow lower right*) has become almost completely filled with sediment by 1947. Photo by Topographic Division, British Columbia Department of Lands and Forests.

may, however, make a precise correlation impossible. Rate of ablation in the vicinity of the glacier tongue shows little correlation with previous winter snowfall, but, except in the 1941 season, it shows a close correlation with fluctuating summer temperatures as measured at New Westminster and Pemberton Meadows. Taylor's observations in August, 1943 (personal communication), indicate that the ice surface of Helm Glacier at the 5,900-foot level was lowered by melting at a rate of as much as 2 inches daily, and a snow surface on a northern exposure at the 5,600-foot level by as much as 3 inches daily. Loss in depth of snow in one 24-hour period (August 19–20) was $3\frac{1}{2}$ inches at the 5,700-foot level and only $\frac{1}{4}$ inch at the 6,800-foot level; the loss in depth of ice in this same period was 1 inch at the 5,900-foot level and $\frac{1}{2}$ inch at the 6,400-foot level.

Observations made by the Dominion Water and Power Bureau on the east tongue of Helm Glacier and on Sentinel Glacier indicate changes in depth of ice in a period of one year (1947–1948) varying from a loss of more than 30 feet at one point to a gain of about 5 feet at another. In spite of wide variations recorded, a general increase in the rate of thinning with the approach to the ice margins or ice front is clearly marked. On Helm Glacier the ratio of loss in depth at points within 100 feet of bare ground to loss in depth at points more than 400 feet from bare ground was almost 2:1, and on Sentinel Glacier this ratio was considerably greater than 2:1.

An explanation of the rapid shrinkage of alpine glaciers within the past few decades has been sought in meteorological records taken at two near-by stations: Pemberton Meadows, 30 miles north of the map-area, where continuous records have been made since 1913, and New Westminster, 30 miles south of the map-area, where continuous records have been made since 1895 and discontinuous records since 1879.[3]

The records of annual precipitation at New Westminster suggest, in spite of marked year-to-year fluctuations, a precipitation cycle with a period of approximately 33 years (fig. 4). As far as can be ascertained, however, the precipitation of the latest cycle is not significantly different from that of its predecessor or, indeed, from that of the period 1859–1863. The records from Pemberton Meadows show in the shorter period a similar trend to that of New Westminster, and it can be expected that the same cyclic change in precipitation has taken place both at the former station and in the intervening mountains of the Mount Garibaldi map-area. Other climatic factors being constant, a period of abnormally great precipitation would lead to increased accumulation of snow in the névé areas of the glaciers and, at some undetermined time later, to an advance of the glacier front. Conversely, a period of reduced precipitation would lead to a reduction in the accumulated snow and, in time, to a retreat of the front. Certain observed fluctuations in the rate of glacier retreat may well stem from past periods of excessive or deficient precipitation, but as yet no evidence is available from the map-area to establish a definite relationship, let alone determine the time lag between any such changes in precipitation and the observed fluctuations of the ice fronts. The rapid ice retreat prevailing since the 1920's cannot, moreover, be correlated with any significant change in recorded precipitation rates in the previous three decades.

[3] Meteorological records made at New Westminster in the period 1859–1863 by the Royal Engineers are also available but are probably not strictly comparable with later observations.

The trend of the prevailing temperatures, as indicated by moving 10-year averages (fig. 4), closely parallels the rate of retreat of Warren Glacier. Very slow retreat of Warren Glacier during the first decade of the twentieth century coincided with a notably cool period. Between 1906 and the mid-1920's the annual rate of glacier retreat increased from about 15 feet to fully 100 feet, corresponding roughly to the period of greatest temperature rise. Since the mid-1920's the rate of retreat of this glacier has changed but little, and, judging from the scale of temperature changes at New Westminster and Pemberton Meadows, the mean temperatures at Warren Glacier have been nearly constant. A perfect parallelism of temperature change and glacier retreat is, however, lacking. The retreat of the Warren Glacier front from a nearly stationary position about the beginning of the century started when some of the lowest summer and lowest annual temperatures in the New Westminster record were still to be attained. Rapid retreat of the west tongue of Helm Glacier was not attained until after the mid-1920's, although by that time relatively warm temperatures had prevailed for several years. Such anomalies may perhaps be accounted for by the delayed effects of previous periods of unusually light or unusually heavy accumulations of snow on névé areas or, alternatively, to periods of excessive or abnormally light melting in these source regions during a succession of warm or cool years.

Other climatic factors may contribute to glacier fluctuations and especially to the minor changes in rates of retreat, but little is yet known of their importance. Long-term changes in wind direction or wind strength could be important in determining changes in the size and shape of glacierets, many of which are nourished by the drifting of snow from adjacent ice-free slopes. Even such large glaciers as Helm may receive important contributions of snow by drifting, and any significant change in winds could have its effect at some later date along the ice front. No evidence is available, however, to demonstrate that such a change has taken place. The duration of bright sunlight during the year may also be important in determining the rate of ablation, but meteorological records at Vancouver taken between 1913 and 1941 indicate no significant change in this factor over a period of years. Changes in the seasonal distribution of precipitation could also have an effect on the glaciers, but again no long-term change has been indicated in the records.

Changes in the rates of glacier flow attributable to changes in the level of the névé surface, to surface gradients of the ice, to shape and area of cross section of the glacier, and to physical characteristics of the ice as determined by surface temperatures, depth, and amount of contained rock debris can all be responsible for changes in the rate of retreat. So, too, the changes in the configuration of the glacier front, the degree of exposure of the ice to sunlight and to heat radiated from near-by rock surfaces, and the protection, if any, afforded by superincumbent morainal debris can be important in influencing glacier recession. With all these factors contributing to fluctuations, it is rather surprising that a fairly close correlation of retreat and temperature is actually found.

DATES OF THE RECENT GLACIAL MAXIMUMS

Forests must once have occupied large parts of the areas exposed by the recent retreats of the glaciers, judging from the numerous relics found. A nunatak in the

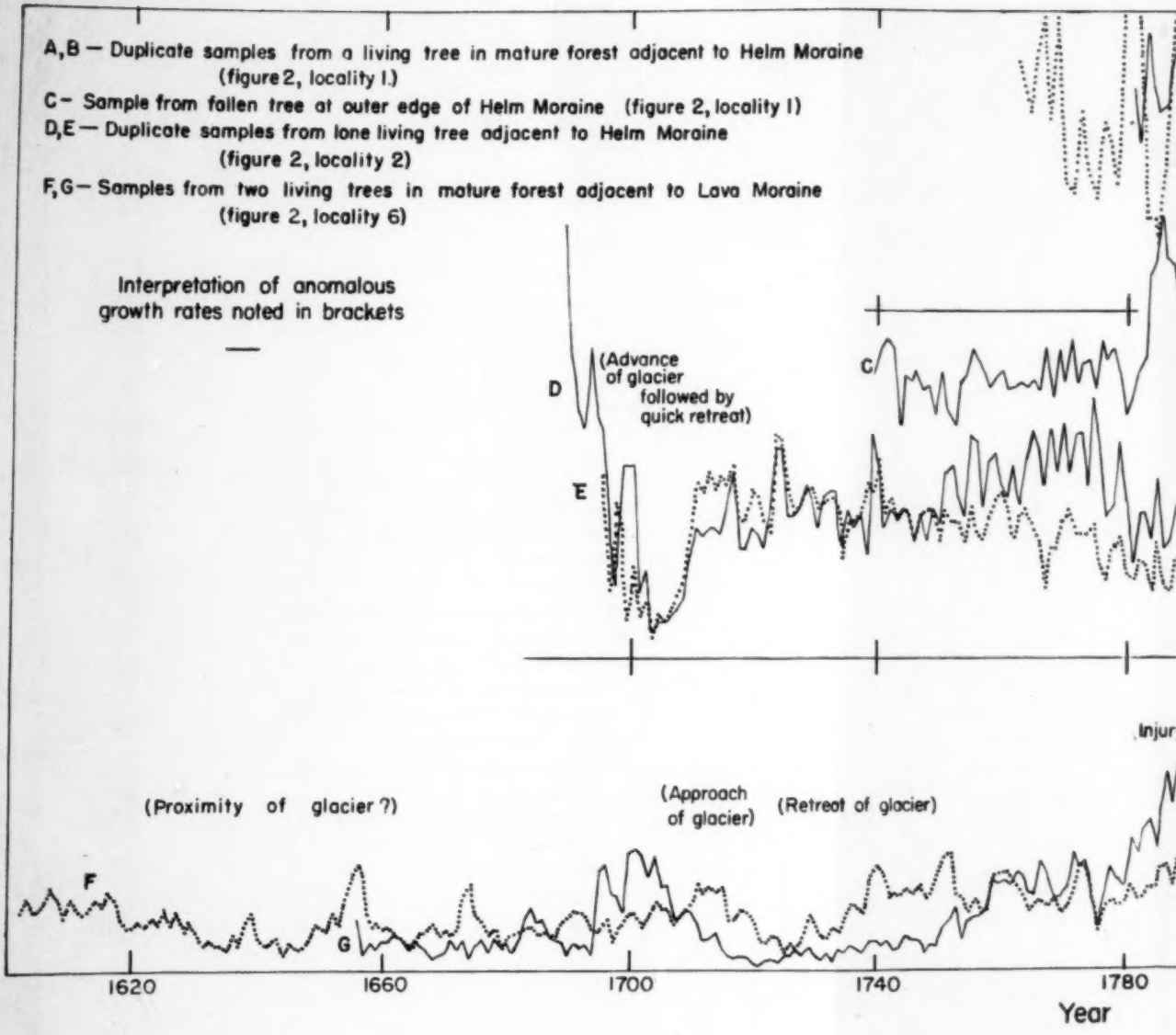
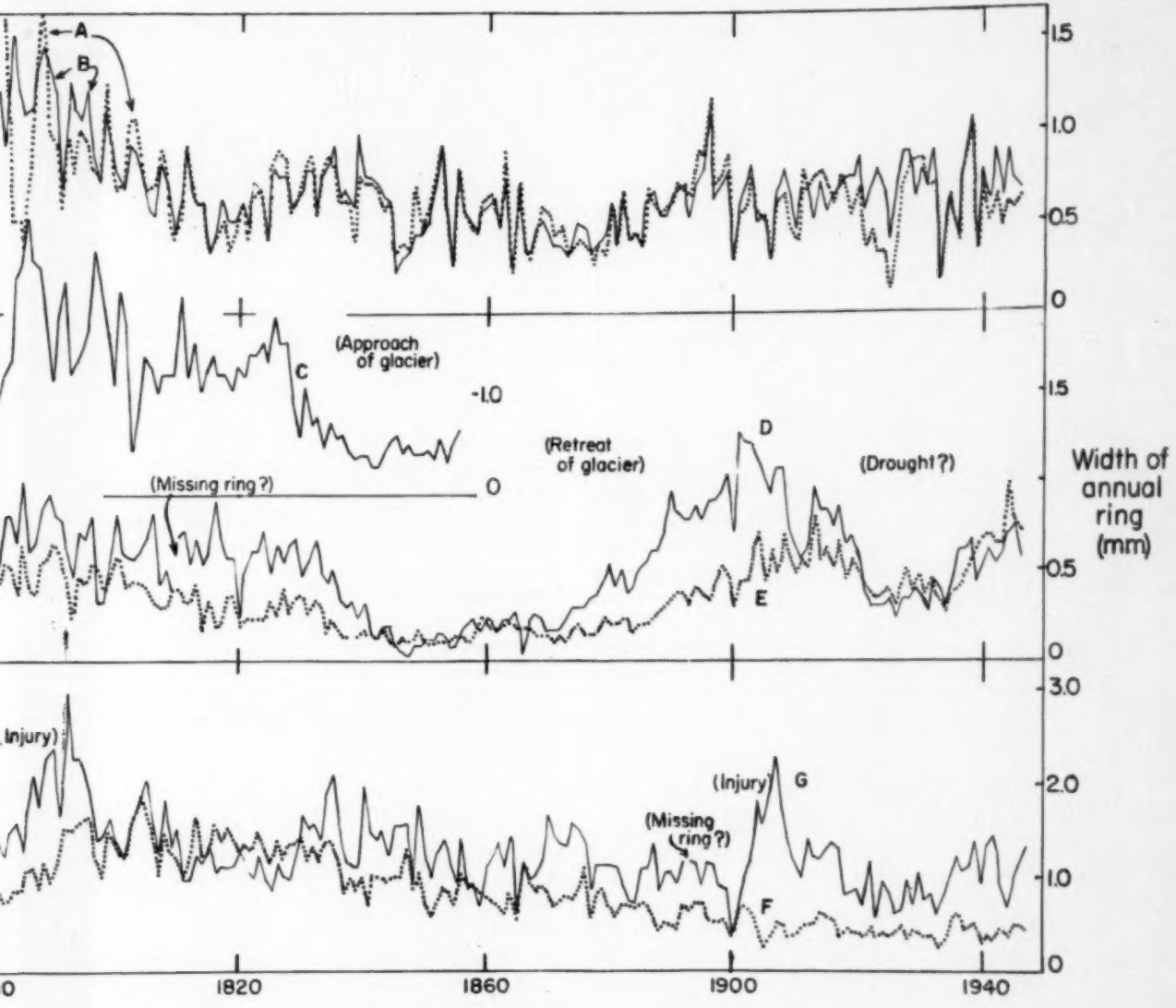

A,B — Duplicate samples from a living tree in mature forest adjacent to Helm Moraine (figure 2, locality 1.)
C — Sample from fallen tree at outer edge of Helm Moraine (figure 2, locality 1)
D,E — Duplicate samples from lone living tree adjacent to Helm Moraine (figure 2, locality 2)
F,G — Samples from two living trees in mature forest adjacent to Lava Moraine (figure 2, locality 6)
Interpretation of anomalous growth rates noted in brackets
(Advance of glacier followed by quick retreat)
D
E
C
Injur
(Proximity of glacier ?)
(Approach of glacier)
(Retreat of glacier)
F
G
1620
1660
1700
1740
1780
Year

FIG. 5.—Measurements of growth rings of tre



of trees near the outer limits of moraines

Garibaldi Névé, 2 miles east of the summit of Mount Garibaldi, covered by ice until probably the early 1940's, bears several tree stumps whose roots still penetrate cracks in the otherwise barren rock. The trees lived at an altitude of 6,100 feet, a level which is at, or perhaps slightly above, present tree line for this vicinity. On the stoss side of a knoll east of The Cinder Cone and exposed since 1940 by retreat of the east tongue of Helm Glacier, other stumps have been found, again rooted in cracks in the rock. In this locality, beneath a thin mantle of fresh ground moraine, an old soil still exists, in which can be found but slightly decomposed relics of sod and of fir branches with needles still attached. Several stumps up to 3 feet in height stand near the northern edge of Sphinx Moraine $\frac{2}{3}$ mile from Garibaldi Lake on the slopes of a lava knoll, and on the eastern or stoss side of this knoll (fig. 2, locality 4) a tree 18 feet in height still remains upright. Photographs show that the stumps and tree were still buried by Sphinx Glacier in 1912, and at the time of the ice maximum they were covered to a depth of between 100 and 200 feet. Presumably because of its position in front of the precipitous face of the lava knoll, the tree was not swept away by the ice, although it may well have been pushed back against the cliff face and held there by the glacier at the time of its flood. It stands today bent and crushed, though still erect and still possessing a few naked branches, testifying to the force of the glacier and to the more favorable conditions for forest growth in the times before the ice advance. It may be assumed from these relics that forests once existed in those areas which were exposed by ice retreat up to at least 1940 and which had altitude, exposure, and soil conditions similar to the existing timber-covered slopes beyond the recent ice limits. With the advance of ice into these old forests, most of the trees would be either broken off near their bases or uprooted and then either buried in moraine or carried forward with the ice and exposed to continual decay. Thus the few scattered relics found today may represent only a small portion of the former forest cover.

One group of uprooted trees, apparently overturned by the final advance of the ice, lies at the edge of the moraine of the east tongue of Helm Glacier (fig. 2, locality 1). These trees fell in a direction away from the former ice tongue and into the mature forest beyond the limits of devastation. There they were partly buried by outwash from the glacier before it receded from its maximum position. Although the bark has long since decayed from the prostrate logs and the outermost inch of the wood is checked and weathered, it is still possible to sample them and to measure the widths of growth rings. A comparison of the rings of one of these logs with those of a living tree in the mature forest a few yards away was attempted; but, although a general similarity of growth rates can be detected if the fallen log is assumed to have been killed within a few years of 1860, no wholly satisfactory correlation could be made (fig. 5). The possibility of missing or false rings in the cores from either tree could readily spoil any perfect correlation, and the presence of such missing or false rings is indicated by studies of duplicate cores from different sectors of the trees (fig. 5, *A* and *B*, *D* and *E*). The absence of a perfect correlation cannot, therefore, be taken as evidence disputing the death of the fallen tree about 1860.

A gradual, though marked, diminution in growth rate in the uprooted tree toward the end of its life is apparently not

matched by a similar diminution in the living tree. A similarly marked, if gradual, diminution in growth is noted in the rings of one of the broken stumps on the north side of Sphinx Moraine (fig. 2, locality 4). In both trees the retardation of growth may be attributed to the gradual approach of the glacier, which seems to have been directly responsible for their untimely ends. That this influence of the glacier on tree growth does not extend far beyond its front is shown, however, by the absence of marked contraction of rings in living trees within a few yards of the ice limits on both Helm Glacier (fig. 2, locality 1) and Warren Glacier (fig. 2, locality 5), as well as in other trees far from any ice. In the hope, however, that similar retardation of growth attributable to proximity of ice might provide a dating of some of the glacial maximums, trees have been cored at several other localities. Samples from two localities have yielded results that may be significant. A lone hemlock (*Tsuga mertensiana*), a few feet from the ice limit of the east tongue of Helm Glacier (fig. 2, locality 2), shows a gradual diminution of growth rate between 1830 and 1845 (fig. 5, *D* and *E*) followed by a gradual return to normal growth rate in the period 1875–1890. These changes may be attributed to a gradual advance of Helm Glacier in the first period and to a gradual retreat in the latter. A sharp reduction in the rate of growth of this tree in the period 1700–1712 may likewise have been brought about by an earlier and more rapid advance of Helm Glacier. The second locality at which a marked, though gradual, retardation of tree growth has been noted is at the limit of Lava Glacier (fig. 2, locality 6). Here two separate trees showed a gradual diminution of growth in the one to two decades immediately prior to 1725 (fig. 5, *F* and *G*), followed by a gradual increase in rate of growth in the succeeding few decades. Other periods of retarded growth in the seventeenth century are noted in the cores from these same two trees, but none can be attributed with any certainty to nearness of ice.

The character of vegetation on moraines of the Mount Garibaldi map-area, as on those of other glacier areas, gives some clue to the period during which the land has been exposed and hence provides a figure for the minimum time since the glacier climax. Scattered individuals of pioneer herbs may take root within a few years of exposure of barren moraine, and various species of trees may take root shortly thereafter. In most places in the map-area, however, two to four decades elapse after the exposure of the moraine surface before species of forest trees are capable of germinating and surviving the seedling stage in any numbers. On a part of the moraine of Lava Glacier exposed for forty years, for example, not more than a few dozen young trees, a few feet in height, have developed per acre. However, once a cover of low vegetation has been established on the moraines, spread of the three principal tree species of this region and altitude—the alder (*Alnus tenuifolia*), the mountain hemlock (*T. mertensiana*), and lovely fir (*Abies amabilis*)—seems to take place rapidly. In some places on the outermost moraines, the alder may be the predominant tree, elsewhere the fir. The low-growing alder and the slow-growing hemlock are soon overtopped by the fir in these youthful forests, and for a time this tree is the dominant species. The alder seems doomed to virtual extinction in the shady environment of such a forest, but the hemlock thrives. Judging from the mature forests beyond the recently glaciated areas, the hemlock, not the fir, becomes

ultimately the dominant species. In most places a progressive change in plant associations following the pattern outlined above is found between the currently retreating glacier fronts and the trimline beyond the outermost moraine, and in most places the largest trees on these moraines are no more than a few decades old. A different situation exists, however, near the outer limits of Sphinx and Lava moraines and possibly, too, at the northwest corner of Warren Moraine. On these, a fairly well-defined boundary can be recognized which separates a sparse, shrubby vegetation, similar to that found on the outermost moraines of the other glaciers of the map-area, from a dense fir-alder forest; and beyond this again is a sharply defined boundary separating the fir-alder forest from a mature hemlock forest beyond the moraines. The oldest tree that could be found in the outermost part of the fir-alder forest in front of Lava Glacier (fig. 2, locality 6) took root shortly prior to 1775. Assuming that a few decades may have elapsed after the ice left this point before re-vegetation could take place, the retreat from the climax of Lava Glacier may have commenced in the first half of the eighteenth century. This agrees well with the postulated glacial climax of 1725 deduced from growth rings in the two trees of the mature forest a few tens of feet beyond this point. The oldest tree found in the young fir-alder forest northwest of Sphinx Moraine (fig. 2, locality 3) took root about 1748, indicating that Sphinx Glacier also began retreating from its climax position in the first half of the eighteenth century. Presumably, the advance and retreat of these two glaciers were in response to regional rather than local conditions, and other glaciers of the map-area may have attained maximums in this same period. If this is true, the advances of these other glaciers in the middle or latter part of the nineteenth century attained or exceeded the earlier limits and destroyed the record of the vegetation.

Still other indications of glacier fluctuations may be found in the records of the tree rings, even though their interpretation may be difficult or uncertain. Many changes in growth rates may be brought about by conditions wholly independent of glacial activity. A marked increase in the width of rings for a period of a decade or more may stem, for example, from mechanical injury caused by wind, avalanches, or rock slides. Other changes might, however, be directly attributable to fluctuations of a near-by glacier. An advance of the glacier which caused the destruction of a part of a forest might leave the surviving trees with a greater exposure to sunlight and might thus change the rate of growth according to the response of the particular species to light. It would be difficult to reconstruct the history of a particular tree from the rate of growth alone, but the events may be deciphered from independent evidence and the dates inferred from the record of the rings.

A change in the level of the water table on the lava flow in the valley west of Warren Glacier (fig. 2, locality 5) at some recent date can, for example, be postulated on the basis of existing evidence. The construction of the outermost moraine of the glacier has blocked and diverted one of its meltwater streams onto the lava flow and through an old forest. Decayed stumps still stand in the new stream channel, testifying to the recency of the diversion. It has not proved possible to correlate the rings of these decayed stumps with those of living trees along the stream bank, but in one stream-side tree a zone of rotten wood may record the drowning of part of its

root system about 1715. If so, Warren Glacier reached its maximum extent, at this outermost moraine, at about the same time as did Lava and Sphinx glaciers.

Mountain glaciers attained a greater size in the eighteenth and nineteenth centuries than at any time since the waning of the Late Wisconsin ice sheet, judging from evidence from Lava, Warren, and Helm moraines.

The recent moraines of Lava Glacier are characterized by boulders of red dacite derived from a lava flow which forms the valley wall west of the glacier. This lava predates the "Ring Creek flow" (p. 359), which, as already noted, was extruded while remnants of the Cordilleran ice were still lingering in the lowland of the southwestern part of the map-area or had very recently disappeared from it. Beyond the 1725± limit of Lava Glacier, erratics of the red dacite are conspicuously absent, and whatever glacial drift is present is attributable to Cordilleran glaciation.

The recent terminal moraine of Warren Glacier is built against the southern lava flow from Clinker Mountain, in places burying the lower part of the trunks of still living trees. No morainal debris exists on the surface of the flow beyond this recent ice limit, even though the lava has apparently been there since the waning stages of the Cordilleran ice sheet.

Similar testimony is obtained from Helm Glacier (fig. 3). The Cinder Cone at the front of this glacier bears on its northern slope and summit scattered ice-borne erratics, of which most are vesicular basalt, many are quartz diorite, but only a few are sedimentary rocks such as exist around and presumably beneath Helm Glacier. Basaltic lava is exposed at the southwestern base of the cone and probably extends eastward for a short distance under Helm Glacier, but by far the greatest extent of this basalt is to the north and northeast. The nearest occurrence of quartz diorite other than as relatively minor dikes and sills is about 2 miles to the north and northeast of The Cinder Cone, where a body many square miles in extent is known to exist. The erratics on the cone thus appear to have been derived from the north or northeast, whence the Cordilleran ice is known to have moved, rather than from the south and southeast, whence Helm Glacier flows today. If so, it follows both that The Cinder Cone was once covered by the Cordilleran ice sheet and that only its southern part, which bears the nineteenth-century moraine of Helm Glacier, has been covered by that glacier since the melting of the ice sheet exposed the surface of the cone.

ADVANCE OF THE GLACIERS TO RECENT MAXIMUMS

The alpine glaciers of the map-area were, as noted, in an advanced state during the eighteenth and nineteenth centuries, and there is some doubtful evidence that at least one, the Lava Glacier (p. 367), may have extended almost to its maximum limit during the seventeenth century. The record of the moraines indicates, too, that at least three of these glaciers had not exceeded these limits in all post-Pleistocene time prior to the eighteenth century. The dead trees exposed by the recent recession of the ice are, moreover, a record of the time when the glaciers were distinctly less extensive than they have been within the last two or three centuries. When, then, did the glaciers advance from this less extensive condition? At least one of the dead trees on Sphinx Glacier (fig. 2, locality 4)

shows a gradual contraction of growth rings at its outer edge attributable to the approach of the ice which finally overwhelmed and killed it; and it was hoped that, by correlating the rings of this dead tree with those of living trees near by, it might be possible to establish the date at which Sphinx Glacier advanced beyond this spot. Such a hope has not been fulfilled, although, with luck, such a correlation and dating might yet be achieved elsewhere. Fortunately, however, delta growth in Helm Lake provides a record of the duration of the extended stage of Helm Glacier and hence makes possible a rough dating of the time of its advance to modern proportions.

The delta in question (fig. 3) is being built by a meltwater stream from the western part of Helm Glacier. This stream, on crossing the western end of The Cinder Cone, picks up pyroclastic debris and deposits it on the flats beyond or in the waters of Helm Lake. As the lava dam forming Helm Lake bears erratics and till left by the Cordilleran ice, Helm Lake must have been in existence since the disappearance of that ice. The topographic relations of Helm Glacier to The Cinder Cone are such that, at any time when Helm Glacier is in an advanced stage, meltwater from the western part of the glacier would flow across the west flank of the cone into Helm Lake and would then contribute to the delta. On the other hand, it seems probable that, at such time or times when Helm Glacier may have been distinctly less advanced than in 1947, meltwater or normal stream runoff from the area now occupied by the western part of the glacier might have escaped around the southern and eastern base of The Cinder Cone, bypassing Helm Lake entirely. If so, the age of the Helm Lake delta is not a measure of all post-Pleistocene time but only of that part of post-Pleistocene time during which Helm Glacier was sufficiently advanced to cause overflow of meltwater across the western end of The Cinder Cone.

An advance of the front of this delta into Helm Lake, averaging 140 ± 20 feet along a 1,200-foot front between 1928 and 1947 is clearly shown in photographs. If it be assumed that the area of the delta has increased at a uniform rate throughout its history, its growth has required only about 375 years. On the same basis a few additional decades may have been required for the construction of a series of small fans built on its surface by meltwater streams issuing from the glacier at its most advanced stages. Adjustments to this estimate of about four centuries for all delta construction may, to be sure, be necessary. The rate of delta growth for the period 1928–1947 may be higher than formerly and the estimated age correspondingly low because (1) the flow of meltwater was abnormally great as a result of rapid glacier recession and (2) the depth of Helm Lake along the advancing delta front may have been less than in the parts previously filled. On the other hand, the early rate of delta growth may have been higher than in recent years because (1) erosion of the cinder which makes up the delta was probably fastest in the earliest stages of stream cutting and (2) a much higher proportion of the cinder was initially carried into the lake than at present, now that an appreciable part of the stream load is being deposited on the upper surface of the delta. In view of the compensating character of these factors, the estimate of the age of the delta is probably in error by no more than two or three centuries, and it is clear that delta growth has occupied only a relatively small part of the 10,000+ years of post-Pleistocene time.

The stage of glacier advance at which overflow of meltwater and delta construction started can be ascertained approximately. The present stream course was initiated across a saddle at an altitude of 5,900 feet in the original surface of the cinder deposit. Water ponded by an advancing Helm Glacier against the southern slope of The Cinder Cone would certainly have overflowed, had it risen to this level; and it is conceivable that the water may have begun its escape by underground percolation before that time. The permafrost within the otherwise permeable cinder, which makes possible the existence of a lake in the summit crater, presumably, however, retarded or prevented underground leakage. In the absence of such leakage the ice front would have had to rise to the 5,900-foot level before overflow could take place or to a somewhat higher level, had the ponded water been able to flow under or through the ice tongue. The ice in 1947 was, however, sufficiently impermeable to maintain a small lake ponded against the south side of The Cinder Cone, even though the ice front to the west rose only a few feet above the water level. In 1947, moreover, the high point on the ice front at the southeastern face of The Cinder Cone had an altitude of 6,000 feet; and, had it not been for the recently developed breach in the western end of the cinder barrier against which the glacier is built, the ice front would have attained a still higher level. It is almost certain, therefore, that an advance of Helm Glacier to a stage comparable with that of 1947 would have been more than adequate to cause overflow of meltwater. It seems safe to conclude, also, that, since the disappearance of the Cordilleran ice from the 5,700-foot level in this vicinity, Helm Glacier has been less advanced than it was in 1947 for all but the four centuries, more or less, representing the life of Helm Lake delta. That Helm Glacier was more extensive than in 1947 for a century or two immediately after the disappearance of Cordilleran ice is not inconceivable; but a single period, of about four centuries' duration, culminating in the climaxes of the early 1700's and mid-1800's, seems more reasonable. This recent advanced condition, stemming apparently from widespread climatic conditions, was almost certainly shared for a comparable period of time by the other glaciers of the map-area.

THE ANOMALOUS BLACK TUSK GLACIER

The history of Black Tusk Glacier differs markedly from that of its neighbors, in that this body of ice was apparently very much larger in late Wisconsin time than it has been within the past few centuries. In its retreat from its latest maximum this glacier has left almost 1 square mile of moraine characterized by black dacite rubble, which is as yet virtually devoid of vegetation. For distances of up to $\frac{3}{4}$ mile beyond the limits of this modern moraine to the east, north, and northwest, within areas now covered by mature alpine or forest vegetation, numerous erratics of the same characteristic black dacite can be found (fig. 2). Striations, apparently of glacial origin, radiating from The Black Tusk, occur at several localities in this same area, and at one place—at the 5,500-foot level northwest of the mountain—such striations cross older, southerly-trending striae attributable to the Cordilleran ice sheet. This, together with the presence of black dacite fragments on the surface of the glacial drift at comparable altitudes, indicates that the great extent of Black Tusk Glacier occurred after the final recession of Cordilleran ice from the 5,500-foot level. However, at an altitude of

4,900 feet in a stream-cut bank 1¼ miles north of The Black Tusk, boulder clay containing numerous black dacite fragments is overlain by a boulder clay in which boulders of greenstone and quartz diorite are predominant among the coarser debris and which is, therefore, of Cordilleran origin. The relationships in this exposure suggest an advance of Cordilleran ice over either Black Tusk ice or Black Tusk moraine, and the absence of a sharp boundary between the two tills favors the former interpretation. If the lower boulder clay is contemporaneous with the erratics and striae at higher levels, it would appear that this maximum extent of Black Tusk Glacier occurred at a very late stage in Wisconsin time, after the ice sheet had receded from the 5,500-foot level but before it had disappeared from the 4,900-foot level. That Black Tusk Glacier was very much larger than now at a time when Helm Glacier, only 2 miles to the east, was almost certainly smaller than it is at present seems inescapable from the evidence available. The black dacite rubble and even the striae might have been attributed to ancient rockslides, had it not been for the presence, 1¼ miles northwest of the mountain, of erratics and striae on the distal slope of a ridge separated from The Black Tusk by a cirque ½ mile in width and several hundred feet in depth. The striae and the distribution of erratics, it seems, can be attributed only to local glaciation of late Pleistocene age.

Two ancient rockslides, beheaded by recent cirque cutting, do exist on the eastern slope of The Black Tusk (fig. 6), and these may provide at least a partial explanation for the small size of the modern Black Tusk Glacier as compared with its late Wisconsin predecessor. These rock streams, though descending almost to the 6,300- and 6,500-foot levels, respectively, well below the upper limit of Cordilleran ice, have been unglaciated and are hence of late or post-Wisconsin age. The debris, composed of black dacite rubble resting on Upper Cretaceous sediments, has clearly been derived from the central spire of The Black Tusk. The source, moreover, was almost certainly in that part of the spire above and to the north of the 7,250-foot contour; for, along the southeast ridge below that level, only dark-red dacite and older sediments are present at a high enough altitude to feed the rockslides. The lower ends of these slides have, however, been isolated from the central spire by headwall recession of a small cirque by a distance amounting to at least 300 feet. Should the debris have originated from the central or northern part of the spire rather than from the southern part, a much greater head-wall recession would be indicated. A marked lowering of the crest of the ridge, amounting in all probability to no less than 200 feet and possibly to as much as several hundred feet, must have accompanied this head-wall recession, judging from the ridge slopes now prevailing. It may well be, therefore, that Black Tusk Glacier has, in late and post-Wisconsin time, by its own erosive activity destroyed much of its own potential gathering ground. No such lowering of ridge crests is noted around the head of Helm Glacier, where rare Cordilleran erratics testify to the preservation of the Wisconsin land surface, or on the summit of Mount Garibaldi, the source of Lava and Warren glaciers, where relics of the original crater walls can still be found. This contrast in the amount of lowering of the cirque head walls may at least help to explain why Black Tusk Glacier alone was smaller during the glacial maximums of the past few centuries than it was in late Wisconsin time.

EARLY ALPINE GLACIATION

A record of pre- or early Wisconsin glaciation is provided by low-level cirques in the southern Coast Mountains. In the southwestern part of the Mount Garibaldi map-area, cirques exist on the north slopes of ridges rising no more than 5,000 feet above sea level, and only 18 miles south of the map-area cirques exist on ridges less than 4,000 feet in altitude. Unlike the cirques at and above the 6,000-foot level which were occupied by ice within the past few centuries or are still so occupied, these low-level cirques contain no morainal ridges to indicate the presence of alpine glaciers at any time since the disappearance of the Cordilleran ice sheet, and, indeed, the floors of some contain undisturbed till laid down by this ice sheet. Many of these basins together with their surrounding ridges, moreover, were completely buried during the more advanced stages of the Cordilleran ice and at that

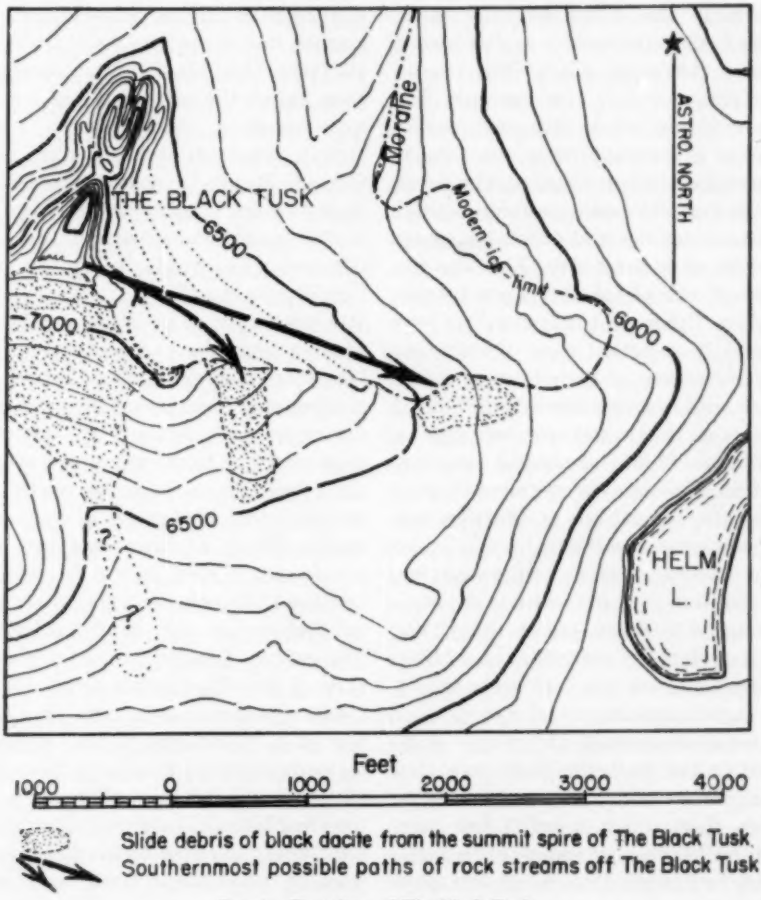


FIG. 6.—East slope of The Black Tusk

time could have contained no alpine glaciers. The alpine glaciation and the low firn line to which these cirques are to be attributed must, therefore, be assigned to some period preceding the latest climax of the Cordilleran ice, that is, to early or pre-Wisconsin time.

COMPARATIVE DATA

The glacial and climatic history of the Mount Garibaldi map-area is by no means unique, inasmuch as elsewhere in western North America and, indeed, throughout the world where detailed studies have been made a similar record is found (Matthes, 1942; Flint, 1945). From many sources, for example, has come evidence of a prolonged period since the Pleistocene when the climate was warmer and drier than it is at present and when glaciers and undrained desert lakes were smaller and vegetation more arid than they are today. This was the period of the so-called "climatic optimum" or "thermal maximum." It was followed by a period when the climate was more severe, either wetter or cooler or both, and when glaciers and undrained lakes were larger and vegetation of a more humid character than at the present time. This was the so-called "period of climatic stress" or, as Matthes has aptly named it, the "little ice age."

The concept that the alpine glaciers were relatively shrunken during the waning stages of the last ice sheet, though not new, is a relatively unfamiliar one. Many workers have, instead, held to the theory that the disappearance of the last ice sheets was accompanied by a slow withdrawal of the formerly much extended alpine glaciers which originally fostered the ice sheet. Such a belief has been shown by Mannerfelt (1943) to be inapplicable in Scandinavia, where the main ice sheet "was dead even when the highest peaks of the ice divide area emerged as nunataks" and was apparently receiving no contributions of ice from cirques, although at least one of these cirques now contains an active glacier. Although the southwestern part of the Cordilleran ice sheet was not completely stagnant at the time of its decline, this ice sheet, like its Scandinavian counterpart, appears to have received no significant contributions from local glaciers. Indeed, in several localities in both the southwestern and the northwestern quadrants of the Cordilleran sheet, local glaciers had shrunk to a size little, if at all, larger than their modern representatives before the waning of the ice sheet was complete. According to Mackin (1941), the local trunk glaciers of the Cascade Range in northwestern Washington "were notably small at the time of the maximum extent of the Puget Glacier [an important distributary of the Cordilleran sheet] and . . . they did not attain large size at any subsequent time." Mackin refers also to evidence suggesting "the interesting possibility that the local glaciers may have been somewhat larger during an *early phase of the last glacial stage* previous to the maximum stand of the Puget Glacier." Farther north in the central Coast Mountains of British Columbia, W. A. D. Munday (1947, 1948) has found, almost up to the fronts of existing alpine glaciers, glacial striae attributable to the southward-moving "Homathko Glacier," another distributary of the Cordilleran sheet. He concludes (personal communication) that not since the waning of the Cordilleran ice exposed their basins have these alpine glaciers extended more than a few hundred feet below their present limits. Hanson (1932) likewise concludes from his work in the Portland Canal area on the northwestern coast of British Columbia

that alpine glaciers were relatively reduced in size before the ice sheet had disappeared. Other evidence on this point is available from the Stikine Valley, which passes through the northern Coast Mountains from the interior of British Columbia to the Alaskan coast. Upstream from the mouth of Scud River, kame terraces and meltwater channels occur on the lower slopes of Stikine Valley at altitudes of little more than 500 feet and were evidently formed along the margins of a dwindling Cordilleran ice tongue extending southwesterly from the interior. The seaward drainage of these low-level channels shows that the three glaciers which today reach the floor of Stikine Valley downstream from Scud River were not sufficiently advanced in late Wisconsin time to have obstructed the flow of meltwater from the ice sheet. Damming of the Stikine drainage above the 500-foot level would almost certainly have occurred had any one of these glaciers, 8–17 miles long today, been 1–3 miles longer and from 500 to 1,000 feet thicker than they are at present. Here, too, it seems that alpine glaciers were little, if at all, larger during the last stages of the Cordilleran ice than they have been in historic time. In the Tuya-Teslin area of the northern interior of British Columbia, Watson and Mathews (1944) concluded from the presence of strand lines of a late glacial lake within the Atsutla Range that deglaciation of this range was almost complete while the adjacent plateau was still buried to a depth of at least 500 feet by the ice sheet. In the last three areas as well as in the Garibaldi area the small size of alpine glaciers in late Wisconsin time suggests that the climate then prevailing was little, if at all, more severe than that of the present. In this connection Hansen (1947) from pollen studies in the Puget Sound area of northwestern Washington states that "the presence of a nearby ice front evidently did not cause a boreal or even a sub-boreal climate," although he does suggest from the distribution of western white-pine pollen through the peat-bog successions that "the climate was cooler and moister during the early Postglacial than at any time since." If, as seems probable, the early retreat of the trunk glaciers of the northern Cascade Range was brought about by climatic causes, a marked moderation, if not a change to post-Pleistocene conditions, had taken place prior to the climax of the Puget Glacier and perhaps of the Cordilleran sheet itself.

Several possible causes, acting alone or in combination, may serve to account for the small size of alpine glaciers during the decline of Cordilleran ice. An abrupt moderation of climate, such as would lead ultimately to marked recession of the sheet, would bring about an early retreat of local glaciers, while distributaries of the sheet, nourished by enormous reserves of ice penned in its encircling ring of mountain ranges, might continue to flow with little diminution in size for many centuries. The ice sheet, moreover, might through its great size maintain a local climate (Brooks, 1926) favoring its perpetuation under conditions in which the smaller alpine glaciers could not survive. Once a general retreat had begun, the ablation would tend to be much more rapid in relatively small alpine glaciers exposed to heat radiated from adjacent ice-free slopes than anywhere on the large unbroken ice and snow surfaces of the sheet itself. Finally, a thick, protective mantle of morainal debris may have been present on the waning distributaries of the ice sheet, whereas only a few of the alpine glaciers heading in areas of rapidly disintegrating

or easily eroded rocks, as do Warren and Black Tusk glaciers today, would have been equally well protected.

The time at which the "little ice age" began, estimated as 4,000 years ago (Matthes, 1942), might appear to differ markedly with the time, four centuries ago, when, for the first time since the Wisconsin, Helm Glacier in the Garibaldi area may have attained its present size. The discrepancy, however, may well lie in the definition of the *beginning* of the "little ice age." In spite of the fact that the change to the prevailing moderately severe climate may have begun forty centuries ago or even earlier, the time when it first attained the severity of the earlier half of the twentieth century may prove a more definite, if not a more convenient, "beginning" of the "little ice age." Matthes based his figure in part on the work of Hanson, who estimated the age of the Bear River Delta in the Portland Canal area and suggested that it had been entirely constructed from glacial debris eroded during the "little ice age." Hanson infers that the rate of delta growth at the mouth of Bear River is a direct function of the extent of glaciers in the Bear River drainage basin, and his estimate of 4,000 years of growth is based on the assumption that the glaciers were of their present size throughout this period. Alternatively, however, the glaciers of that area may have been a fraction of their present size for a period much longer than the estimated 4,000 years and equal to, or greater than, their present size for a correspondingly shorter period. The varved clays on the floor of former Tide Lake (Hanson, 1933), also in the Portland Canal area, may provide one of the best records of the "little ice age," for this lake came into being only with the development of an ice dam following the advance of a glacier a few centuries or millenniums ago to a position approximating that of the 1920's. Hanson was able to measure only the uppermost varves of the resulting lacustrine sediments and was required to estimate on very limited evidence the total depth of the deposit. Direct measurements from borings might yield a reasonably accurate figure on the duration of the lake and of the glacial conditions fostering it, and such a figure could be one of the more reliable guides to the duration of the severe climate in the "little ice age." A correlation of annual rings in "fossil" trees exposed by ice retreat with those of old, but still living, trees may some day yield independent data on the duration of the "little ice age."

The time of the latest, if not the greatest, climax of the "little ice age" in the Cordilleran region is fairly well defined, being so recent that its record is clearly preserved in vegetation, living and dead. The forests that have sprung up since this glacial maximum are still immature, especially by comparison with the climax forests beyond the glaciated areas. In many places the new growth has attained the state of the fir-alder forest which exists in front of Lava Glacier; elsewhere the new vegetation is still less mature. Trees killed during the final advances of the ice, as before Helm Glacier, are not yet decayed beyond recognition. Trees drowned in two moraine-dammed lakes during the latest advances of Scimitar and Klinaklin glaciers (Munday, 1936, 1937), 150 miles northwest of Mount Garibaldi, were still standing in lake waters in the 1930's. How long a tree could remain erect under such circumstances is not well known, but many trees similarly drowned shortly before 1860 in two slide-dammed lakes near Garibaldi were still standing eighty-seven years later, whereas only the

stumps are left of trees drowned by meltwater from Warren Glacier in 1715(?) (p. 368). Scimitar and Klinaklin glaciers probably attained their recent climaxes not more than about two centuries ago. Franklin Glacier, in this same area, probably attained its climax not less than a century ago, judging from the presence in the early 1930's of a tree "possibly 60 or 70 years old" on its modern moraine (Munday, 1936). Alsek River, draining through the St. Elias Range, was recently dammed by the extension of Lowell Glacier across its course (Kindle, 1947). The lake thus formed extended for many miles upstream along both Alsek and Dezedeash rivers, and, though the obstruction has since disappeared, the strand lines of the former lake remain. McConnell (1904) concluded from the undecayed character of driftwood on the lake shores, from the immature state of the reforestation on the lake floor, and from legends of the Indians that the lake may have existed within the century prior to 1904. It is of interest that the driftwood, which was described as "undecayed" by McConnell in 1904, is referred to as "fairly well rotted" by Kindle in 1947.

Earlier climaxes of the mountain glaciers have been reported. McConnell, for example, stated that the higher of two Dezedeash shore lines, which he also attributed to an ice-dammed lake, was covered by "ordinary forest growth of the region," and he suggested that this earlier lake dated back some hundreds of years. It is possible, if not probable, however, that the higher shore line should be assigned to the late Wisconsin rather than to the early part of the "little ice age." Kerr (1936) from his studies in Craig Valley, near Stikine River, concluded that glaciers there reached a maximum post-Pleistocene extension, some 25 miles beyond their present limits, about 1,000 years ago. The deposits of terraced drift that he attributed to this period may, however, like similar features in many of the other valleys of the eastern Coast Mountains, be attributed to the late Wisconsin Cordilleran ice. Munday (1936) reports an old till sheet extending downvalley from the limit of modern moraines of Franklin Glacier, but he adds that it is oxidized to a depth of about a foot and correlates it with the Wisconsin till occurring in the vicinity of Vancouver. The glacial maximums of the last three centuries have undoubtedly been exceeded in times gone by, but that these earlier maximums should be assigned to the early part of the "little ice age" seems doubtful. If the limit of about four centuries to the more severe part of the "little ice age" applies not merely for Helm Glacier but throughout the Cordilleran area, glacial maximums of the early part of this period should be clearly recorded in the vegetation, except where that record has been destroyed by later and greater advances.

Not all glaciers, even within the same general area, reach their maximums simultaneously. Some, such as Muir Glacier in southeastern Alaska (Cooper, 1939), are believed to have reached their maximums about the end of the eighteenth century; others, such as Harriman Glacier in southern Alaska (Cooper, 1942), were advancing into maturely vegetated areas in the 1930's. Within relatively restricted areas and especially among smaller alpine glaciers, the maximums seem to be approximately coincident. It is noteworthy that the two maximums of the Mount Garibaldi map-area have their counterparts elsewhere in the Cordillera. Lawrence (1949) concludes from botanical studies that the maximum extent of the glaciers of Mount Hood, a

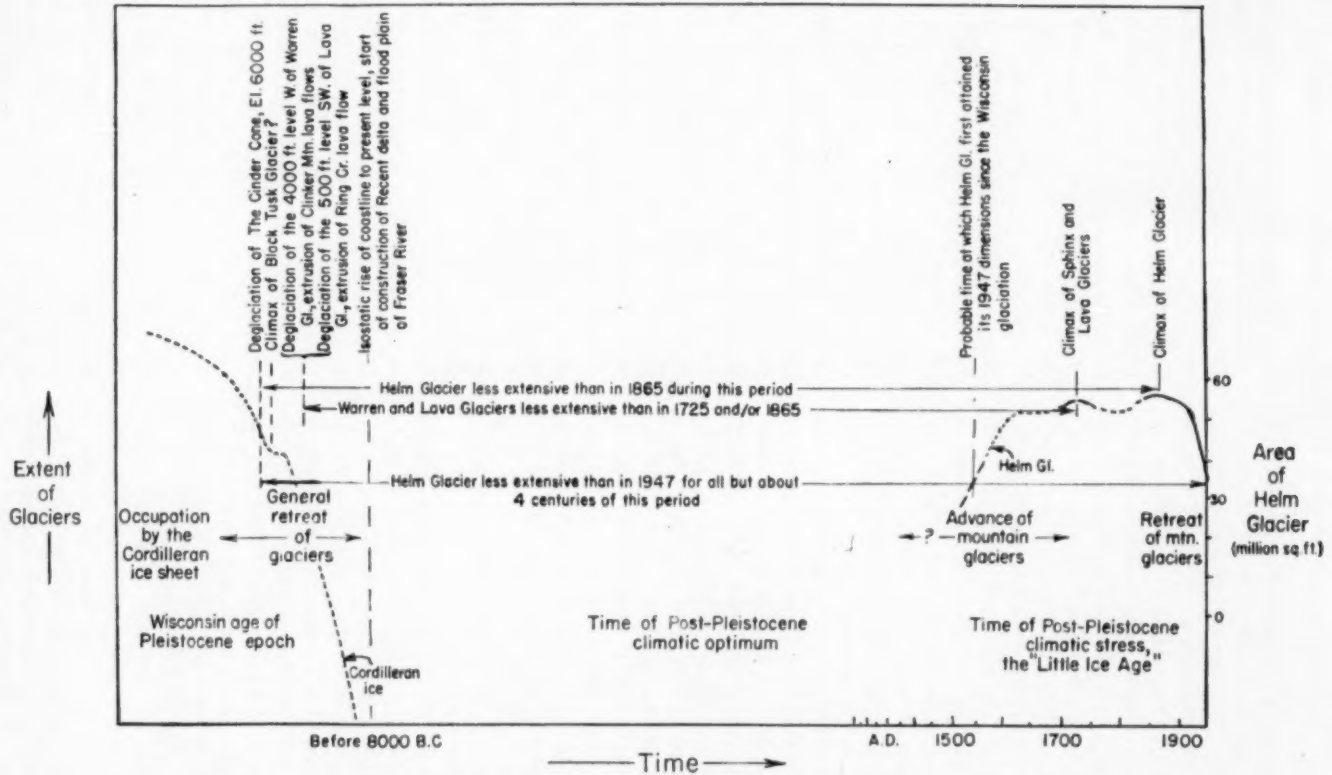

Deglaciation of The Cinder Cone, El. 6000 ft.
Climax of Black Tusk Glacier?
Deglaciation of the 4000 ft. level W. of Warren Gl., extrusion of Clinker Mtn. lava flows
Deglaciation of the 500 ft. level SW. of Lava Gl., extrusion of Ring Cr. lava flow
Isostatic rise of coastline to present level, start of construction of Recent delta and flood plain of Fraser River
Probable time at which Helm Gl. first attained its 1947 dimensions since the Wisconsin glaciation
Climax of Sphinx and Lava Glaciers
Climax of Helm Glacier
Helm Glacier less extensive than in 1865 during this period
Warren and Lava Glaciers less extensive than in 1725 and/or 1865
Helm Glacier less extensive than in 1947 for all but about 4 centuries of this period
Helm Gl.
Extent of Glaciers
Occupation by the Cordilleran ice sheet
General retreat of glaciers
? — mountain glaciers
Advance of
Retreat of mtn. glaciers
Wisconsin age of Pleistocene epoch
Time of Post-Pleistocene climatic optimum
Time of Post-Pleistocene climatic stress, the "Little Ice Age"
Cordilleran ice
Area of Helm Glacier (million sq ft)
60
30
0
Before 8000 B.C
A.D.
1500
1700
1900
Time

FIG. 7.—Extent of ice and glacial history in the Mount Garibaldi map-area in late and post-Wisconsin time

little more than 300 miles south of Mount Garibaldi, was attained about 1740, and he notes evidence of a halt or minor advance at about 1840. The Illicillewaet Glacier of the Canadian Selkirks (Sherzer, 1907), less than 300 miles northeast of Mount Garibaldi, was just beginning its latest prolonged retreat in 1887. An immature alder forest, extending for several hundred feet beyond the 1887 limit, testifies to an earlier and greater extent of this glacier, probably within the previous one to two centuries. Living trees close to the edge of this immature forest testify, in turn, that Illicillewaet Glacier had never exceeded this outer limit in the preceding five centuries. A climax in 1894 of Saskatchewan Glacier, 60 miles north of the Illicillewaet, is reported by Field (1949), who was able to establish that this climax had not been exceeded in the previous 140 years at least. Observations made by Cooper (1916) in 1914 in the Mount Robson region, 100 miles northwest of Saskatchewan Glacier, indicate that the recession of Robson Glacier from its outermost moraine had been continuing with several interruptions for at least 116 years. Since the largest tree on the youngest moraine at that date was only 13 years old, a halt in the ice retreat a few years or decades prior to 1900 is indicated. Thus, throughout the width of the Cordilleran area, we find the record of two climaxes of the glaciers, one at some time within the eighteenth century, the other within the nineteenth century, earlier by a few decades perhaps in the west than in the east, but nonetheless a correlative climax.

The more recent glacial history of the Mount Garibaldi map-area conforms, within limits, to that of a considerable part of the Cordillera, judging from the meager evidence now available. It might be reasonable to suppose, therefore, that the earlier glacial history (fig. 7) of the Mount Garibaldi map-area itself provides some clues to the earlier glacial history of the larger area.[4]

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THE SYSTEM Al_2O_3-H_2O[1]

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ABSTRACT

A study has been made of phase equilibria in the system Al_2O_3-H_2O. Results are expressed in the form of a pressure-temperature diagram, showing regions of stability for the crystalline phases gibbsite, boehmite, diaspore, and corundum. The diagram shows pressure and temperature minima (2,000 lb/in², 275° C.) for the stable existence of diaspore, and the significance of these data for the natural formation of diaspore is discussed.

INTRODUCTION

Metamorphic rocks may be viewed as an inscription telling of the conditions existing in various parts of the earth's crust throughout geologic time. A common type of metamorphic rock originated at the earth's surface as an accumulation of weathering products—aggregations of low-temperature minerals composed chiefly of SiO_2, Al_2O_3, MgO, iron oxide, alkalies, H_2O, and CO_2. With burial and consequent increase in temperature and pressure, the original sedimentary rock was converted into a metamorphic rock as the mixture of crystal structures responded to the changing conditions. Changes in environment were sufficiently slow that a close approach to equilibrium among the phases was realized as pressure and temperature became increasingly greater. The direction of pressure and temperature change was then reversed after a time, and these metamorphosed rocks were brought to the surface for our study. Fortunately, during this reversal the record was not wiped out. To read this record we need a key, part of which is supplied by field and laboratory studies of metamorphic rocks, which have made possible the rating of certain metamorphic mineral assemblages with regard to the relative severity of pressure-temperature conditions imposed. But we lack "fixed points" on this complex pressure-temperature scale, as well as other types of information.

A few fixed points are available from laboratory studies, as, for example, those concerning phase assemblages in the system MgO-SiO_2-H_2O for which a pressure-temperature diagram has been derived (Bowen and Tuttle, 1949). This same type of information is needed for other systems, one of the most important being the system Al_2O_3-SiO_2-H_2O. It was as part of a study of this ternary system that an investigation of phase equilibria in the binary system Al_2O_3-H_2O was made. Equilibrium relations among compounds in this binary system can be applied directly in the consideration of equilibria in the ternary system. Within the system Al_2O_3-H_2O are found the compounds gibbsite ($Al_2O_3 \cdot 3H_2O$), boehmite ($Al_2O_3 \cdot H_2O$), diaspore ($Al_2O_3 \cdot H_2O$), corundum (Al_2O_3), bayerite ($Al_2O_3 \cdot 3H_2O$), and gamma alumina (Al_2O_3), of which the first four are known as minerals.

Earlier investigations of the system are summarized by Fricke (1944). These have consisted chiefly of thermal dehydration at atmospheric or lower pressure but were in every case unsuccessful as equilibrium studies because of the slow-

[1] Manuscript received June 1, 1950.

[2] Norton Company, Worcester, Mass.

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ness of reaction. One recent equilibrium study under hydrothermal conditions has been described (Laubengayer and Weisz, 1943) which, however, was limited with respect to the pressure available.

The problem of attaining equilibrium among the phases is a very serious one in a laboratory study of the system Al_2O_3-H_2O as well as in the system Al_2O_3-SiO_2-H_2O. It has not been possible, for example, to reverse the reaction boehmite → diaspore, and consequently equilibrium in this case was not approached from the high-temperature side. The consistency of the data, using different starting materials, is the chief reason for believing that in its broad features the equilibrium diagram presented for the system Al_2O_3-H_2O is correct.

METHOD OF INVESTIGATION

EQUIPMENT

Pressure vessels used were of the Morey-Ingerson type (Morey and Ingerson, 1937), as modified by Morey for connection to a pressure pump. The pump delivered water pressure through steel tubing to the pressure vessels, which were heated in tubular resistance furnaces equipped with temperature regulators. Temperature was measured with chromel-alumel thermocouples inserted in wells at the top and bottom of the pressure vessels. Pressure was measured directly on a Bourdon gauge and could be varied independently of the temperature. The error in temperature measurement was less than $\pm 10°$ C., while pressure measurements were reliable to ± 50 lb/in^2 in the range up to 4,000 lb/in^2 and to ± 400 lb/in^2 in the high-pressure range.

PROCEDURE

Raw materials and mixtures.—Raw materials used for most of the runs were precipitated alumina gel and γ-Al_2O_3. The gel was made by mixing aluminum nitrate solution with ammonium hydroxide in the cold. Some samples were highly purified by electrodialysis; others were prepared simply by decantation washing. Gamma alumina samples were purchased,[4] having a stated purity of 99.9+ per cent Al_2O_3 with a fineness of less than 0.1 μ. One batch showed some corundum in its X-ray pattern, but another showed only diffuse γ-Al_2O_3 lines. Other raw materials included gibbsite[5] ("alumina scale" from Bayer precipitation tanks), bayerite made in the laboratory by carbonation of sodium aluminate solution, and crystalline boehmite made by hydrothermal treatment of gel.

Seed crystals were ordinarily added to the raw materials in making equilibrium runs, the most important of these being diaspore. A sample of the natural mineral[6] from Chester, Massachusetts, was used for this purpose. This specimen of diaspore appeared to be very pure on microscopic and X-ray examination. The gibbsite, bayerite, and boehmite described above were also used as seed crystals, as well as corundum.

Two types of mixtures were most commonly used. One was of γ-Al_2O_3 plus seeds and the other of alumina gel plus seeds. Samples weighing 10–20 mg. were wrapped in envelopes of platinum foil, and heated in the pressure vessels. Because of the large heat capacity of these vessels, the time required to bring a sample up to temperature was appreciable. To minimize the heating-up period, the general practice was to bring the furnace to red heat before inserting the pressure vessel. In this way the temperature of the vessel reached the operating temperature in 1–2 hours. This time was

[4] Linde Air Products Company.

[5] Supplied by Reynolds Metals Company.

[6] Specimen obtained from the U.S. National Museum, No. 1733.

short by comparison with the time at operating temperature, which was usually 2 days or longer. At the end of a run the vessel was quenched in cold water, room temperature being reached in 2–3 minutes.

Obtaining equilibrium.—There is only one rigorous criterion for equilibrium in this type of investigation, which may be stated as follows: If, while pressure and temperature remain constant, the crystals of one phase grow while crystals of a second phase disappear, then the first phase is stable with respect to the second. In accordance with this criterion, the proper method for obtaining the data needed for locating the diaspore-corundum transition curve, for example, would be to use mixtures of diaspore and corundum and find a temperature above which (at constant pressure) the diaspore all disappeared and converted to corundum and below which the corundum all disappeared and converted to diaspore. Finding this temperature for a series of pressures would complete the determination of the transition curve. It was found that, while the conversion of diaspore to corundum is carried out quite readily, the reverse reaction fails to take place within a reasonable time and was not realized in these experiments. In other words, corundum persists as a metastable phase much below its transition temperature. Therefore, this straightforward method fails, and the criterion set up for equilibrium is too rigorous for practical application in studies of this system.

An indirect method was therefore compulsory, the method chosen being the use of γ-Al_2O_3 and gel as already mentioned. These provide a reactive or soluble source of Al_2O_3, which will dissolve and recrystallize on seed crystals of the stable phase. The possibility of metastable growth under these conditions must be kept in mind, and this was the reason for using both γ-Al_2O_3 and gel simultaneously in separate envelopes. The attainment of equilibrium is indicated if these two starting materials, which are different structurally, yield the same crystalline phase. But the evidence obtained in this way is indirect and thus not so reliable as would be the direct transformation from one crystalline form to another and the reverse. Confidence in the method rests mainly on the consistency of the results.

With γ-Al_2O_3 and gel, crystallization is rapid, considerable crystal growth taking place in a matter of hours. However, phase equilibrium as evidenced by production of a homogeneous phase was often slow; in particular, a fine-grained type of boehmite commonly formed metastably and was very slow in transforming to diaspore or corundum. Many runs were made in which some of this material persisted, but it was not considered necessary to prolong them because the approach to equilibrium had been indicated by the partial growth which had already taken place, and a few long runs were made which confirmed the fact that the fine-grained boehmite would disappear if given enough time. The times allowed for equilibrium for runs above 250° C. were 2–4 days, and in most cases longer runs caused no further change. Runs of this duration were also used in cases in which equilibrium was not reached but in which the direction of equilibrium was clearly indicated. In the neighborhood of 100° C., runs were prolonged for 2–4 weeks, and homogeneous phases were still not obtained; but here again the direction of equilibrium was indicated.

IDENTIFICATION OF THE PHASES

After quenching, the samples were dried at 110° C., then examined under

the petrographic microscope, under the electron microscope where this seemed advisable, and by X-ray techniques. Many runs were made in which growth of lack of growth of diaspore on seeds was the significant observation. This observation could be made very sensitively because the seed fragments had rough edges exhibiting irregular fracture. Even slight growth showed up clearly as a development of crystal faces on the seeds or as acicular growth starting from the seeds.

TABLE 1

PROPERTIES OF THE COMPOUNDS

Name	Composition	Crystal System	Refractive Index	Optic Sign	Density	Hardness	Source of Data
Corundum	Al_2O_3	Hexagonal	$\epsilon = 1.760$, $\omega = 1.768$	−	4.0	9	Palache, Berman, and Frondel, 1944
γ-Al_2O_3	Al_2O_3	Cubic	$n_D = 1.696$	.....	3.4	.....	Fricke and Hüttig, 1944
Diaspore	$Al_2O_3 \cdot H_2O$	Orthorhombic	$\alpha = 1.702$, $\beta = 1.722$, $\gamma = 1.750$	+	3.4	6.5–7	Palache, Berman, and Frondel, 1944
Boehmite	$Al_2O_3 \cdot H_2O$	Orthorhombic	$\alpha = 1.646$, $\gamma = 1.661$	−	3.02	3.5–4	Bohnstedt-Kupletskaya and Vlodavetz, 1945
Gibbsite	$Al_2O_3 \cdot 3H_2O$	Monoclinic	$\alpha = 1.568$, $\beta = 1.568$, $\gamma = 1.587$	+	2.4	3	Palache, Berman, and Frondel, 1944
Bayerite	$Al_2O_3 \cdot 3H_2O$	Hexagonal	$n = 1.583$	....	2.5	.....	Montero, 1942

PROPERTIES OF THE COMPOUNDS

In table 1 physical and optical properties of the compounds are shown with a reference to the source of the data. Because corundum, diaspore, and gibbsite are found in nature as moderately large crystals, their properties are well established. The properties of boehmite listed in the table are those given recently (Bohnstedt-Kupletskaya and Vlodavetz, 1945) for a well-crystallized specimen from the Urals. Previous data have been less complete, and the refractive indices and birefringence reported were substantially lower. The optical properties of boehmite as determined on large synthetic crystals grown in this investigation are in good agreement with those shown in table 1. The data for γ-Al_2O_3 and bayerite are incomplete and uncertain because crystals of sufficient size have not been available.

PLATE 1

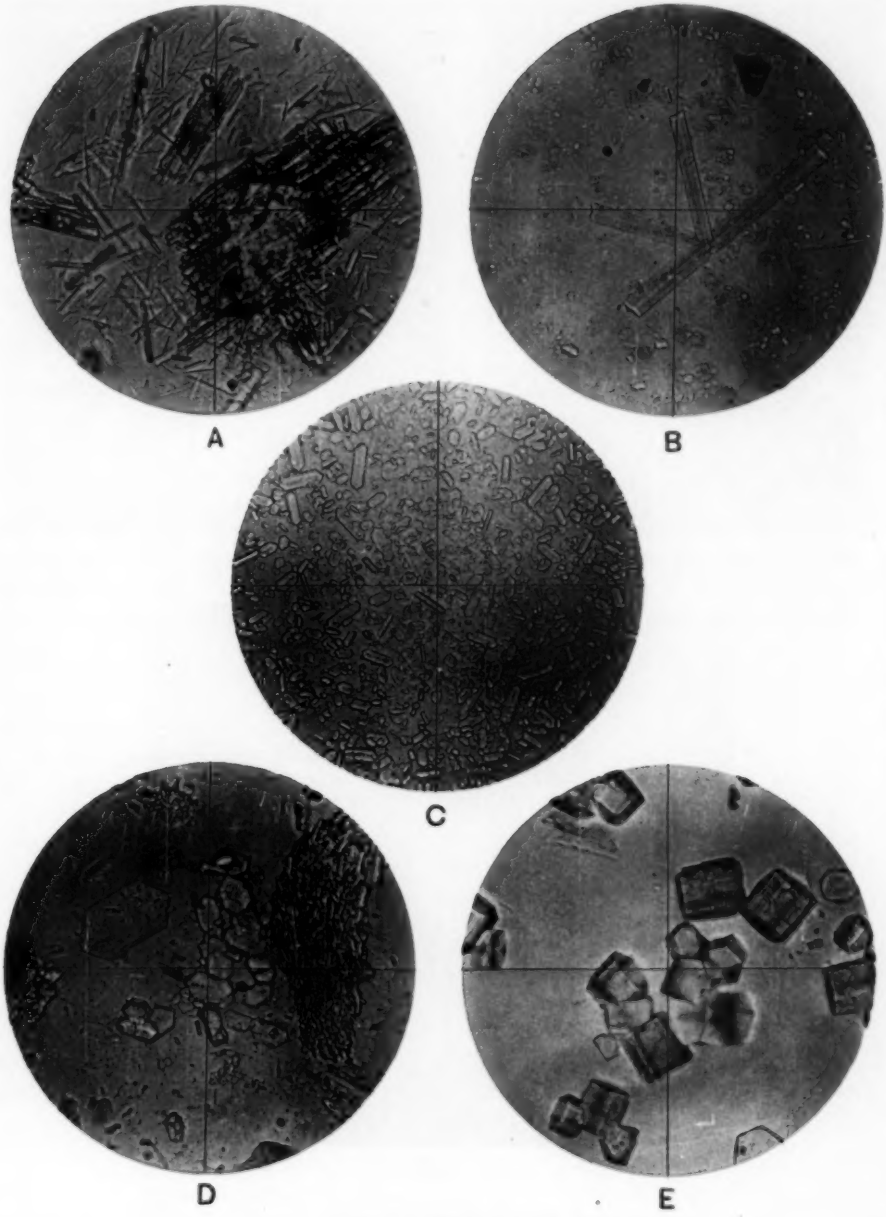
A, Photomicrograph of diaspore grown from seeds. Note triangular seed in central part of diaspore crystal. 313×.

B, Photomicrograph showing twinned diaspore prism grown at high pressure in the absence of diaspore seeds. 313×.

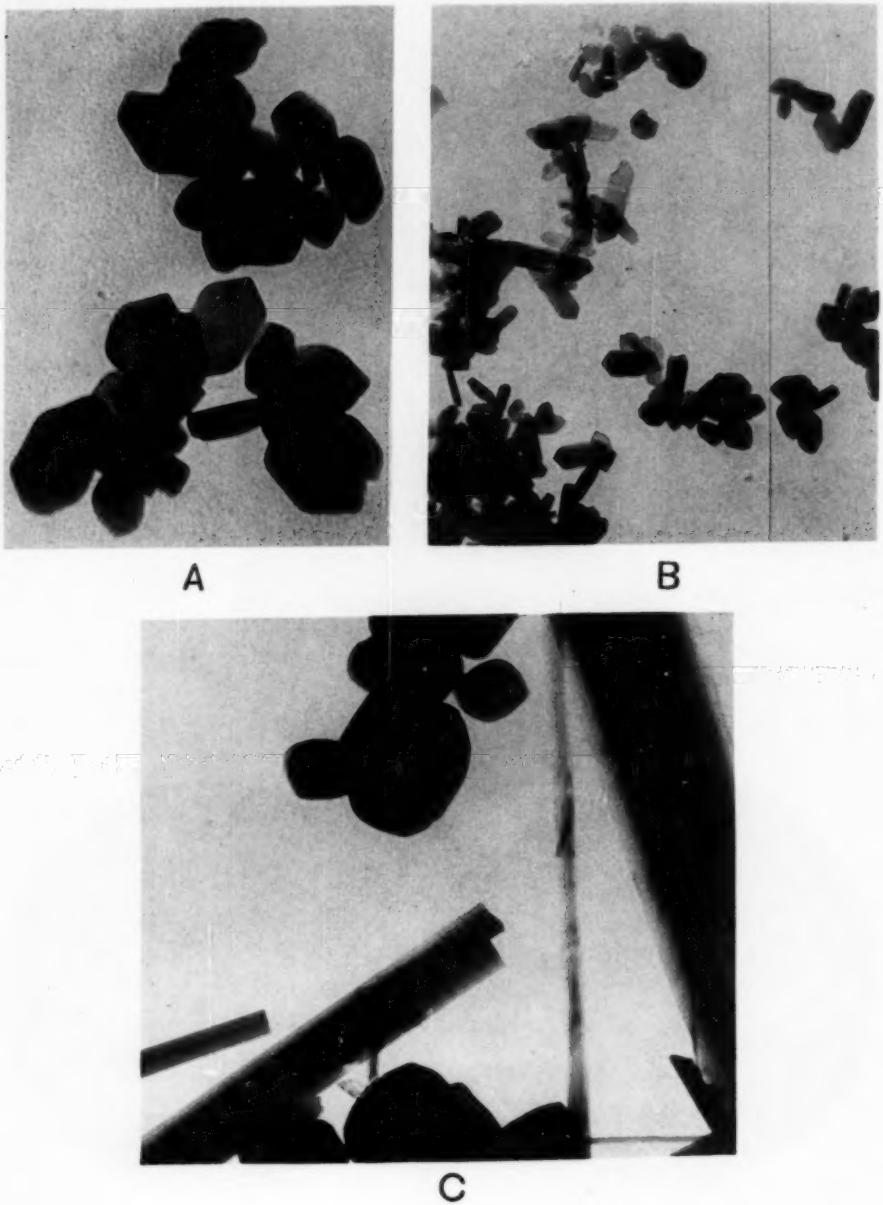
C, Photomicrograph of boehmite crystals. 313×.

D, Photomicrograph of largest boehmite crystals grown. 313×.

E, Photomicrograph showing hexagonal prisms of corundum. 69×.



Photomicrographs of diaspore, boehmite, and corundum



Electron micrographs of boehmite and diaspore

Hydrothermal conditions are especially favorable for crystal growth, making it possible to observe some novel features. Diaspore crystals grew rapidly if seed crystals were present. Crystals of diaspore grown from seeds are shown in the photomicrograph of plate 1, *A*, the large crystal having grown from a triangular seed visible in the center of the crystal. Other crystals in the field grew from smaller seed fragments. In one experiment at very high pressure (table 2) diaspore crystallized in the absence of seeds. These crystals developed as square prisms, sometimes twinned, as shown in the photomicrograph of plate 1, *B*. Growth was predominantly in the *c*-axis direction, commonly resulting in long needles or even fibers. A group of these may be seen in plate 2, *C*, an electron micrograph of a sample containing both diaspore and boehmite. The fiber running across the photograph is only 0.05 μ thick.

Boehmite is normally submicroscopic even when prepared hydrothermally, and this applies also to natural specimens of the mineral. No description of individual crystals of synthetic boehmite is to be found in the literature, and only recently has there been published an adequate description of single crystals of natural boehmite (Bohnstedt-Kupleskaya and Vlodavetz, 1945). Boehmite prepared hydrothermally during this investigation typically appeared isotropic under the microscope and too small to be resolved. Sharp X-ray patterns were obtained, however, and electron micrographs showed well-formed crystals as thin six-sided plates and laths (pl. 2, *A*, *B*).

In a few experiments large boehmite crystals were grown. Plate 1, *C* and *D*, are photomicrographs of two of these preparations. The habit of the crystals is similar to that of the much finer boehmite shown in the electron micrographs. The large crystal in plate 1, *D*, is 150 μ long as compared with 0.5 μ for those in plate 2, *B*. Refractive indices for sodium light measured on these large boehmite crystals are as follows: $\alpha = 1.649 \pm 0.001$, $\beta = 1.649 \pm 0.002$, $\gamma = 1.665 \pm 0.002$. The crystals are biaxial negative and length-fast. These data are in good agreement with those reported recently for natural boehmite (Bohnstedt-Kupletskaya and Vlodavetz, 1945).

Corundum crystallized readily as hexagonal plates, dipyramids, prisms, and needles. Plate 1, *E*, is a photomicrograph of a preparation consisting of hexagonal prisms.

PHASE-EQUILIBRIUM RELATIONS

THE PRESSURE-TEMPERATURE DIAGRAM

Collecting the data on which the equilibrium diagram is based required 250 hydrothermal runs involving 600 separate samples. On the basis of these data, which are summarized in table 2, a pressure-temperature equilibrium diagram for the system was drawn (fig. 1), using a log scale for the pressure axis. Each dot refers to a run listed in table 2. The stability field of each phase is labeled, and the univariant curves (*heavy lines*) forming the boundaries of these stability fields are thus transition curves

PLATE 2

A, Electron micrograph of boehmite crystals. 16,000×.
B, Electron micrograph showing thin boehmite plates. 5,000×.
C, Electron micrograph showing laths and fibers of diaspore mixed with equant boehmite crystals. 5,000×.

TABLE 2

SUMMARY OF DATA FOR THE SYSTEM Al_2O_3-H_2O

HIGH-TEMPERATURE REGION

Water Pressure (Lb/In²)	Temperature (° C.)	Time (Hours)	Crystalline Phases Present*	
			γ-Al_2O_3†	Gel†
Boehmite-Diaspore				
1,500	285	120	bhm	bhm
1,500	310	93	bhm	bhm
2,900	290	101	bhm	bhm
2,000	300	98	bhm, di	bhm + (di)‡
2,000	315	90	di	di
2,000	335	48	di	bhm + (di)
2,000	375	90	bhm	bhm
2,200	340	71	bhm	bhm
2,200	345	98	di	di
2,300	330	192	di	di
2,500	330	67	di	di
2,500	345	98	di	di
5,000	335	67	bhm, di	bhm, di
3,300§	300	76	bhm	bhm
3,400§	310	45	di	di
3,400§	315	48	di	di
3,400§	330	72	di	di
3,400§	340	50	di	di
3,400§	350	76	di	di
3,500	290	117	di	bhm
4,700§	310	132	di	bhm
5,000§	350	50	di	di
6,000	280	93	di, bhm	di, bhm
6,000§	285	37	di	bhm
6,000	295	95	di, bhm	di, bhm
6,000	305	92	di + (bhm)	di + (bhm)
10,000§	255	80	bhm	bhm
10,000§	290	72	di	di
10,000§	295	111	di	di, bhm
10,000§	320	60	di	bhm
15,000§	285	98	bhm	bhm
15,000§	300	98	di	di
15,000§	300	114	bhm	bhm
15,000§	330	45	di	di
15,000§	345	99	di	di
18,000§	315	70	di	di
20,000§	260	62	bhm	bhm
20,000§	275	38	bhm	bhm
20,000§	285	94	bhm	bhm
20,000	285	79	bhm	bhm
Boehmite-Corundum				
500§	385	44	cor	cor, bhm
450§	440	47	cor	cor
1,000§	380	120	bhm	bhm
1,100	345	53	bhm	bhm
1,500	345	192	bhm	bhm
1,500	430	134	cor	cor
2,000§	350	48	bhm	bhm
2,000§	395	118	cor	cor
Diaspore-Corundum				
2,600§	355	136	di	bhm, di
2,600§	360	47	bhm	bhm + (di)
3,200	355	18 days	di	di
3,200§	365	264	cor, di	bhm, di
3,200§	375	47	cor, di	bhm
3,400§	375	40	di, bhm	di, bhm
3,400§	400	48	cor, bhm	di, cor, bhm
3,700§	390	67	di, cor	di, bhm
3,800	345	74	di	di
4,000§	375	69	di, cor	di, cor
4,000	385	13 days	di	di
4,000	395	174	di	di
4,000§	415	48	cor	di, cor
6,000§	380	93	di, bhm	di, bhm
6,000§	405	46	cor	di, cor, bhm
6,000§	415	51	cor	cor
6,800§	440	64	.....	cor
10,000§	390	48	di + (cor)	di
10,000§	405	96	di	di, bhm
10,000§	410	93	di, cor	di, bhm
10,000§	415	72	cor	cor, di, bhm
10,000§	420	74	cor	di, cor
15,000§	380	120	.....	di
15,000§	390	82	cor, di	cor, di
15,000§	390	119	di, cor	bhm, di
15,000§	395	40	di	.....
14,000§	395	64	di	bhm, di
15,000§	395	44	di	di, bhm
15,000§	395	10	di, bhm	di, bhm
15,000§	400	177	cor, di	cor, di
15,600§	415	64	cor	cor
16,000§	415	3	cor, di	di, cor, bhm
17,000§	415	54	cor	di
16,000§	420	70	cor	di, cor
15,000§	425	44	cor	bhm, di, cor
16,800§	425	12	cor	di, cor, bhm
20,000§	375	51	di	di
20,000§	410	92	di, cor	di, bhm
20,000§	410	91	cor	di, cor
21,000§	420	48	cor	di, cor
27,000§	400	51	di, bhm	di, bhm
50,000§‖	390	100	.....	di, bhm

* Abbreviations used: bhm = boehmite, di = diaspore, cor = corundum, gibs = gibbsite, bay = bayerite, gamma = γ-Al_2O_3.

† In virtually all runs two charges were used, one of γ-Al_2O_3 and the other of alumina gel, both containing a very small amount of "seeds." The last two columns of the table refer to the phases present, using these two starting materials.

‡ Parentheses around a phase in the product indicate that very little growth took place on seeds.

§ Fine-grained diaspore crystals were the seeds in runs marked with §. The other runs, unless otherwise indicated, were seeded with diaspore, corundum, and boehmite crystals. Seeds present in the final product are not considered as "Crystalline Phases Present."

‖ No seeds present; diaspore grew without seeds. The pressure, estimated from the degree of filling of the vessel (0.95), using extrapolated values of the vapor-density measurements of Niewenburg and Blumendal (1932) was greater than 50,000 lb/in².

LOW-TEMPERATURE REGION

Water Pressure (Lb/In²)	Temperature (° C.)	Time (Days)	Starting Material*	Crystalline Phases Present
12.4	95	8	gamma†	gibs
12.4	95	8	gel†	gibs
20.8	110	6	gel	gibs
20.8	110	6	gamma	bhm
20.8	110	6	bay	bay
28.8	120	7	gamma†	bhm, gibs
28.8	120	7	gel†	bhm, gibs
52	140	6	gel	bhm
52	140	6	gamma	bhm
52	140	8	gibs	bhm
52	140	8	bay	bay
20,000	125	23	gamma	bhm
20,000	125	23	gel	bhm
20,000	125	23	gibs	gibs
14,000	140	4	gel	bhm
14,000	140	4	bay	bay

* Gibbsite, bayerite, and boehmite seeds present in all mixtures in the low-temperature region. † Alkali added.

along which the two neighboring crystalline phases are in equilibrium with a fluid phase, the latter being liquid above the vaporization curve and vapor below it.

A notable feature of the diagram is the shape of the diaspore field, which does not extend to low pressures. Below a limiting pressure, therefore, the boehmite and corundum fields are adjacent. Accordingly, there is a quadruple point in the diagram where boehmite, diaspore, and corundum are in equilibrium with fluid. This point is also the junction of the three boundary curves—boehmite-diaspore, diaspore-corundum, and boehmite-corundum. The pressure at the quadruple point is 2,000 lb/in² and the temperature is 385° ± 15° C.

This "bottom" for the diaspore field might be expected from a consideration of the relative densities of boehmite and diaspore and applying LeChatelier's principle. Boehmite has a density of about 3.0 and diaspore of about 3.4. Pressure therefore should favor conversion to diaspore, and a negative slope would be expected for the boehmite-diaspore curve. On the other hand, the diaspore-corundum curve, corresponding to a dehydration reaction, should have a positive slope, even though the density of corundum (4.0) is greater

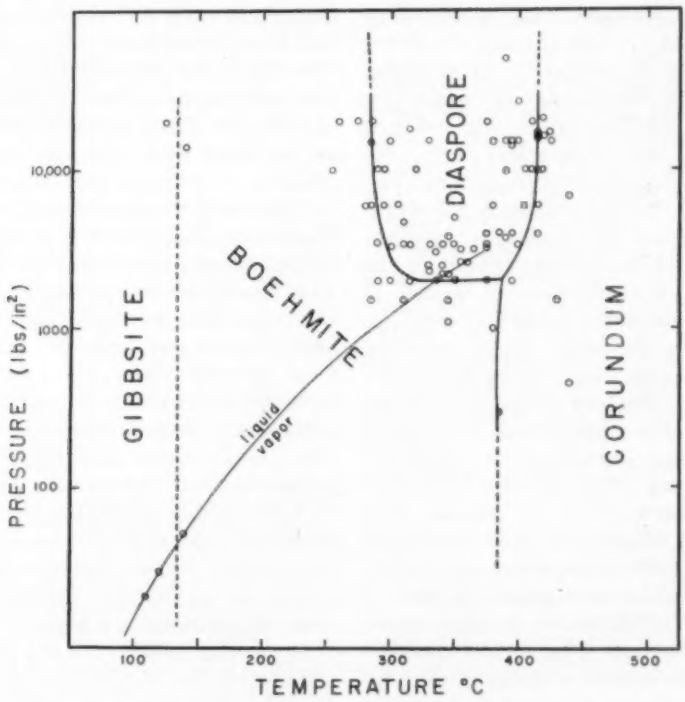


FIG. 1.—Equilibrium diagram for the system Al_2O_3-H_2O. Circles refer to runs listed in table 2. Heavy lines delineate regions of stability of the crystalline phases. Light line is the vaporization curve for water; and, in view of the low solubility of alumina in water, this curve approximates the vaporization curve for solutions in this system.

than that of diaspore, because water vapor is a product of the reaction. Water vapor has a large volume by comparison with the crystalline phases, and pressure should drive the reaction in the direction of the low-temperature phase. Thus the two curves would be expected to have opposite slopes and might intersect. Actually, on the basis of the data obtained, very little slope for the diaspore-corundum curve can be shown. The boehmite-diaspore curve as drawn is a reasonable interpretation of the data. Several points below 2,000 lb/in² between 275° and 385° (table 2) suggest that diaspore is not stable much below this pressure at any temperature.

For runs located near a boundary curve, data are not entirely consistent because of the difficulty in attaining equilibrium. As mentioned earlier, evidence for equilibrium is only indirect, and it is least satisfactory along the boehmite-diaspore transition curve. The reason for this is that there is only negative evidence for boehmite stability, namely, failure of diaspore to grow. In the absence of diaspore seeds and at moderate pressures, boehmite would appear to be the stable phase over the whole diaspore field, but even in the presence of diaspore seeds some conversion to boehmite takes place, with subsequent slow conversion to diaspore. Furthermore, using gel or γ-Al_2O_3 as a starting material, it is possible that growth of diaspore on seeds occurs at temperatures and pressures slightly below those along the transition curve.

Along the diaspore-corundum curve the situation is better, but here concordant results were not obtained with γ-Al_2O_3 and with gel. In the temperature region just to the right of the curve as drawn, γ-Al_2O_3 samples converted to corundum, but gel samples converted to diaspore plus a small quantity of corundum. But, in addition, the diaspore seeds in the γ-Al_2O_3 samples dissolved, sometimes entirely, or in shorter runs some seeds were found in the process of solution, their former jagged edges rounded. This is positive proof that corundum is stable with respect to diaspore under the conditions of those runs. Alongside these, in separate platinum envelopes containing gel instead of γ-Al_2O_3, diaspore crystallized extensively on seeds instead of dissolving. This must then be a case of metastable growth of a phase above its transition temperature. This situation prevailed for about 20° above the transition curve as drawn, and then both starting materials yielded corundum. At temperatures below those of the transition curve, both yielded diaspore. Presumably, if the runs in the transition interval had been continued long enough, the diaspore which grew from gel would all have converted to corundum.

When only the general outline of the diagram is considered, it is seen that the entire area above 380°–415° C. is corundum, whereas below 130° C. it is gibbsite; the area between 130° C. and 275°–300° C. and above 2,000 lb/in² is boehmite, whereas below 2,000 lb/in² the boehmite area extends to the corundum field; finally, between 275°–300° C. and 385°–415° C., above 2,000 lb/in² there is a diaspore area. The data are consistent with these broad features of the diagram, and proof of equilibrium is conclusive, subject to the limitation that approach to equilibrium has been only from the low-temperature phase or from a metastable phase. Having accepted this picture, the transition curves cannot lie very differently from the way they have been drawn. Their precise determination by direct equilibrium data has been found very difficult.

No regions were found for the

stable existence of bayerite ($Al_2O_3 \cdot 3H_2O$) or γ-Al_2O_3. The minerals kayserite ($Al_2O_3 \cdot H_2O$) and tanatarite ($Al_2O_3 \cdot H_2O$) are apparently identical with diaspore. A specimen of kayserite[7] from Cerro Redondo, Uruguay, has the same optical properties and X-ray diffraction pattern as diaspore.

The only previous equilibrium study of the system (Laubengayer and Weisz, 1943) was considerably more restricted than the present one because of equipment limitations. Because Laubengayer and Weisz worked in closed bombs and used a constant degree of filling, they did not attempt to plot a pressure-temperature diagram. On the whole, the present results are in good agreement with theirs. Their suggested boehmite-diaspore transition of 280° C. checks well with our value. Their diaspore-corundum transition point of 450° C. is high; but this was determined at an unknown and very high pressure, at least 50,000 lb/in² (erroneously calculated as 10,500 lb/in²). Their gibbsite-boehmite point at 155° C. is also higher than ours by 25°. This point is very difficult to determine because of the slowness of reaction,[8] and our study of this region was not thorough.

[7] Specimen obtained from the U.S. National Museum, No. 96496.

[8] The difficulties met and the uncertainties still existing in the determination of equilibrium relations in this system add interest to the possibility of working out this type of system from the thermodynamic properties of the crystalline phases and of water. In most cases inability to make the thermodynamic calculations involving concentrated solutions or melts renders such treatment impracticable, but here the solubility of Al_2O_3 in H_2O is so slight that the aqueous phase can be treated as pure water. The free energy of formation of water as a function of temperature is known, and data are available which permit calculation of the free-energy equation for corundum. But, for the crystalline hydrates, the high-temperature and low-temperature specific heats and heats of formation are yet to be determined.

EFFECT OF ADDED COMPONENTS

Addition of any third component to Al_2O_3-H_2O converts it to a three-component system with additional solid phases and more complex relationships. But certain qualitative generalizations can be made, without further data, regarding the effect of adding more components in a quantity small enough that no new solid phases form and assuming that the components do not enter the structure of any of the crystalline phases. The added component, with these restrictions, will be present in the fluid phase, and its chief effect will be to lower the partial pressure of water in the fluid phase. Consider a reaction such as

$$Al_2O_3 \cdot H_2O \text{ (diaspore)} \rightleftharpoons Al_2O_3 \text{ (corundum)} + H_2O \text{ vapor}.$$

This reaction is in the equilibrium state when the decomposition pressure of diaspore equals the partial pressure of water in the fluid phase. If at constant total pressure this partial pressure is lowered (as by an added component), then the temperature of decomposition of diaspore is lowered. However, as long as the added components have low solubilities, only slight lowering of decomposition temperatures would be expected.

Furthermore, it may be seen that such additions would have no effect on the reaction

$$Al_2O_3 \cdot H_2O \text{ (boehmite)} \rightleftharpoons Al_2O_3 \cdot H_2O \text{ (diaspore)}.$$

Water vapor does not enter into this reaction, and therefore the partial pressure of water in the fluid phase will not affect the equilibrium temperature for the reaction. (The *total* pressure does have an effect, given by the Clapeyron equation.) The fact that the lower temperature limit of diaspore stability is not lowered by the presence of other constituents unless there is solid solution is significant with regard to the origin of diaspore in

rocks where other constituents are present. This significance is discussed in the section which follows.

GEOLOGICAL SIGNIFICANCE OF DATA

It is of interest to consider what meaning these data for the system Al_2O_3-H_2O may have regarding the origin of the minerals concerned. The minerals gibbsite, boehmite, and corundum present no special problems. Regarding diaspore, however, these results are in conflict with published theories of the origin of certain field occurrences of diaspore and with the opinion of some geologists and petrologists who have studied certain diaspore deposits. The occurrence of diaspore and the environment during formation will therefore be discussed. Diaspore occurs in three principal types of deposits: (*a*) hydrothermal veins and metamorphic rocks, (*b*) bauxite deposits, and (*c*) nodular diaspore clays. One or more of the other alumina minerals is usually associated with diaspore in these deposits.

In the foregoing discussion it was shown that additional components should theoretically have no effect on the minimum pressure and temperature limits for diaspore stability. Therefore, in this respect the equilibrium diagram can be applied directly to field occurrences, and the most important conclusion about diaspore origin is that, if the diagram is correct and if diaspore crystallized as a stable phase, diaspore must have formed under hydrothermal conditions, that is, at a temperature above about 275° C. and a (steam) pressure of at least 2,000 lb/in², or 130 atm. The first type of diaspore occurrence meets these requirements. In this type of deposit, diaspore is usually associated with corundum, the latter frequently being the major alumina mineral of the deposit. The vein deposits, an example of which is the one at Chester, Massachusetts, are presumably of hydrothermal origin. Also, in the "crystalline rocks" high temperature and pressure were very likely involved. All that need be noted regarding the implication of the phase diagram for these cases is that the temperature was probably moderate. Corundum itself could have formed at high temperatures, but the occurrence of diaspore with corundum suggests a 400°–450° C. range. The possibility exists that diaspore may have formed later than corundum in the 300°–400° C. range, provided that the pressure was sufficiently high. The large, well-developed crystals of corundum found in some deposits could have grown at temperatures as low as 410° C. in the presence of water vapor. The largest and purest crystals of diaspore come from vein deposits, the Chester deposit having excellent specimens.

Bauxite deposits constitute the most abundant source of the alumina minerals and are the only important commercial source of alumina. A typical bauxite contains 55–65 per cent Al_2O_3, 2–10 per cent SiO_2, 2–20 per cent Fe_2O_3, 1–3 per cent TiO_2, and 10–30 per cent combined water (Bateman, 1942). Diaspore occurs as a minor constituent of a few bauxites, the alumina being predominantly in the form of gibbsite or boehmite, sometimes both. The American bauxites are chiefly gibbsite, while those of France and middle Europe are largely boehmite. The geology and petrography of bauxites are well known, but their mode of formation is a subject about which there has been considerable controversy (Edwards, Frary, and Jeffries, 1930). The weight of opinion now favors an origin involving weathering of aluminous rocks *in situ* during prolonged periods of tropical

conditions. Alkalies and then silica were thereby leached out, leaving the alumina and iron as the least soluble. Other modes of formation which have been proposed and which may be important in certain cases include chemical sedimentation (Rozhkova, 1937), secretion of alumina by plants or animals (Bishopp, 1937), and hydrothermal solutions (Lotti, 1940; Vajna, 1948).

In discussing the origin of bauxite, little attention has been paid by most authors to which hydrates are present but only to how alumina hydrates were formed from silicate rocks. A notable exception to this are the reports of Jacques de Lapparent, who has studied French bauxites and others extensively and has maintained consistently (De Lapparent, 1930, 1935, 1946; De Lapparent and Hocart, 1938) that diaspore is found only in those bauxites which were depressed and subjected to high temperatures. Diaspore occurs only in boehmitic bauxites, and it is significant that, in the rare cases in which microscopic crystals of boehmite have been reported (De Lapparent, 1930, 1935, 1946), diaspore has always been present. This is consistent with the present laboratory finding that microscopic crystals of boehmite form only at elevated temperature and pressure, the same conditions necessary for diaspore formation.

A summary of bauxite geology bearing on this question of diaspore genesis was pubished recently by E. C. Harder (1949). His survey indicates that boehmitic and diasporic deposits are of early age, whereas more recent bauxite deposits are of the gibbsite type. Thus boehmite and diaspore occur in those deposits which have had the best opportunity to be covered by an appreciable thickness of later sediments, and these minerals are never found in bauxites formed in the recent past or now in the process of formation, which could not have been covered and subjected to elevated temperature and pressure. As examples of early-formed bauxites, which typically contain boehmite and some diaspore, and of diasporic clays, Harder mentions the following: the bauxites of the Urals and Tikhvin; the diaspore and boehmite clays of southwestern China, Pennsylvania, and Missouri—all of Paleozoic age; the bauxites of Ariège, France, of southeastern France, and of central Italy, Greece, and Hungary—all of Mesozoic age; and the bauxites of Istria and Kashmir—of Eocene age. The bauxites of southeastern United States are also of Eocene age but are the trihydrate type. Bauxite deposits of late Tertiary and Pleistocene age are all of the trihydrate type and comprise those of British Guiana, Dutch Guiana, the West Indies, northern Ireland, Oregon, and parts of India, Africa, and Brazil. Harder makes the following statement at the conclusion of his paper: "General field evidence seems to indicate that first gibbsite forms during the lateritization of aluminum silicates and that aging, pressure and possibly heat change the gibbsite to boehmite and eventually the latter to diaspore."

Diaspore clays are not nearly so widespread, nor have they been so intensively studied, as bauxite deposits. Descriptions of the Missouri and the Clearfield, Pennsylvania, deposits are available. Both are similar in that the diaspore occurs as nodules in a clay of about 70 per cent Al_2O_3, which is associated with flint clays. H. S. McQueen (1939) and V. T. Allen (1935) have published descriptions of the Missouri deposits with theories of their origin. They are agreed that the diaspore clays were formed from the

flint clays by leaching, which dissolved silica and left the alumina behind. The leaching took place chiefly in caverns or sinkholes in dolomitic limestone into which the flint clays had slumped. Circulation of leaching solutions was aided by the associated porous sandstone. R. M. Foose (1944) suggests a similar origin for the Clearfield deposits. These theories of origin furnish an explanation of how the clay was transformed into alumina hydrate but do not suggest any reason for the occurrence in the form of diaspore instead of gibbsite or boehmite. Allen (1935) states that gibbsite accompanies the diaspore, and he regards gibbsite as an intermediate in the production of diaspore from clay but does not suggest what change in conditions would cause the transition from gibbsite to diaspore.

On the basis of the equilibrium diagram of figure 1 and a consideration of the other types of occurrence of diaspore, it would appear that the gibbsite or boehmite was converted to diaspore as a result of the application of heat and pressure. The diagram indicates a minimum pressure of about 2,000 lb/in^2, corresponding to a depth of burial of about $\frac{1}{2}$ km., for the stable crystallization of diaspore. It seems reasonable that diaspore deposits such as those in western Pennsylvania were buried as deep as that. The only question, then, is temperature. If burial of these aluminous deposits has not been sufficiently deep that temperatures of the order of 275° C. were reached as a consequence of a normal geothermal gradient, then there is still the possibility of penetration by hydrothermal fluids or the close approach of igneous rocks to supply the required heat. The occurrence in western Pennsylvania of peridotite dikes cutting the coal measures (Honess and Graeber, 1926; Sosman, 1938) is suggestive in this connection.

Diaspore might conceivably form in nature under conditions of temperature and pressure where it is not a stable phase, as, for example, at low pressures and at temperatures below 275° C. But this is very unlikely in view of the diaspore-boehmite density relationship and the ease with which gibbsite and boehmite form in this range, coupled with the difficulty of forming diaspore even under conditions at which it is stable. Furthermore, if boehmite is present, this cannot be converted into metastable diaspore. Diaspore can form from boehmite only if the pressure and temperature are increased to the point that diaspore is the stable phase. If the structure of diaspore contains sufficient quantities of other ions in the Al^{3+} positions, such as Fe^{3+} or Mn^{3+}, then this structure may be stable at lower pressures and temperatures than pure diaspore, but probably only a trace of other ions occurs in the diaspore of bauxite and diaspore clays.[9]

CONCLUSION

The system Al_2O_3-H_2O was studied for the purpose of finding the equilibrium

[9] Runs were made with diaspore from Clarion County, Pa., as well as from Chester, Mass., as seeds in a mixed ferric hydroxide–alumina gel, to see if a ferrian diaspore might grow and to see if it might grow outside the diaspore field shown in fig. 1. Diaspore grew only within the field shown in fig. 1, and the X-ray pattern of this diaspore was identical to that of diaspore grown from materials free of iron.

Chemical analyses of diaspore from both Clarion and Clearfield counties, Pa., were supplied through the courtesy of the Harbison-Walker Refractories Company. The diaspore samples from the two localities were very similar in composition, containing approximately 3 per cent SiO_2, 2.5 per cent TiO_2, and 0.8 per cent iron oxide determined as Fe_2O_3, besides alumina and water. A microscopic examination of the diaspore, made at the Hays Laboratory of the above company, revealed the presence of siderite, possibly in sufficient quantities to account for the iron present.

relationships among the crystalline phases: corundum, diaspore, boehmite, and gibbsite. A pressure-temperature diagram representing a reasonable interpretation of the data is shown as figure 1. Rigorous proof of equilibrium could not always be obtained because the failure of diaspore and corundum to convert to low-temperature phases prevented approach to equilibrium from both sides. Approach to equilibrium was therefore made by way of two different types of active and metastable starting materials, γ-Al_2O_3 and alumina gel, seeded with the crystalline phases. Because of the consistency of the results from both of these substances, it is considered that in its broad features the diagram represents equilibrium relations.

The shape of the region within which diaspore is the stable phase is the point of chief interest in the equilibrium diagram. An invariant point at which corundum, diaspore, boehmite, and fluid coexist occurs at $385° \pm 15°$ C. at a pressure of $2{,}000 \pm 200$ lb/in^2. Diaspore does not therefore exist as a stable phase at pressures less than this, and its minimum temperature of stability is approximately 275° C. These lower limits apply not only to pure alumina and water but to mixtures of these with other substances. If diaspore does not form in nature as a metastable phase, its presence therefore indicates moderately high-temperature conditions. It is unlikely that diaspore occurring in the clay and bauxite deposits formed metastably, in view of the ease with which boehmite crystallizes even within the region of stability of diaspore.

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Lower Pennsylvanian fossiliferous concretions

GEOLOGICAL NOTES

CONCRETIONS FROM IOWA LIKE THOSE OF MAZON CREEK, ILLINOIS[1]

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During the summer of 1948 Mrs. Ethel Donnelly, of the Burlington Public Schools, brought to our attention some fossiliferous nodular concretions from southeastern Iowa. These are indeed reminiscent of those which are so well known from the vicinity of Mazon Creek in northeastern Illinois. They came from a few feet of Pennsylvanian shale exposed along a ravine in the southwestern corner of the Roy Meyers farm about $\frac{3}{4}$ mile south of the abandoned Oak Ridge schoolhouse, near Dunreath, some 18 miles southeast of Des Moines. We visited the locality in company with Mr. Carl Campbell, of Knoxville, and largely through his continued efforts have secured from it about a hundred concretions which bear plant remains. Furthermore, Mr. Henry E. Efnor, of Newton, and his son have gathered some two hundred similar concretions there. Mrs. Donnelly stated that some years ago one was found which contained a fish, but we have not seen it.

In 1894 A. C. Spencer noted the occurrence of these concretions in southeastern Iowa, and his brief discussion of them was quoted by B. L. Miller in 1901. Nevertheless, the discovery seems to have received scant attention, probably because the locality is given as "north of Dunreath," whereas actually it is some 2 miles west of there and only about 1 mile north. Of course, the possibility exists that Spencer's material came from a different place than ours, but he states that "careful search for similar concretions in the gullies of neighboring streams has not been successful"; and, as far as we have been able to ascertain, no one is aware of more than a single noduliferous locality in the vicinity.

The concretions are similar in size, shape, and composition to those which are so abundant at the classical Mazon Creek area near Coal City, Illinois. Mr. Frederick O. Thompson, of Des Moines, has generously supplied us with extensive collections of excellent material from the latter place, and we are illustrating a few specimens from each locality to demonstrate pictorially their similarity.

All together we have been able to identify the following forms from the Iowa concretions: *Calamites* sp. (one of which is over a meter in length), *Annularia galioides* Lindley and Hutton, *Annularia* sp., *Mariopteris mazoniana* Lesquereux, *Neuropteris flexantea* Janssen, *N. rarinervis* Bunbury, *N. scheuchzeri* Hoffman, and *Odontopteris subcuneata* Bunbury. Every one of these and numerous others are known from the Mazon Creek region, where extensive strip-pit mining of coal has made it practical to secure many times the number of concretions that are brought to light by the erosive work of one creek in the relatively small Iowa area. All

[1] Manuscript received October 14, 1950.

PLATE 1

Fossiliferous concretions from Lower Pennsylvanian shales near Dunreath, Iowa (figs. 1, 3, 6, 7) and Mazon Creek, Illinois (figs. 2, 4, 5) all $\times \frac{3}{10}$. Photographs retouched by Howard Webster. Figs. 1, 2, *Neuropteris scheuchzeri* Hoffman. Figs. 3, 4, *Neuropteris flexantea* Janssen. Figs. 5, 6, *Annularia galioides* Lindley and Hutton. Fig. 7, *Annularia* sp. and *Neuropteris scheuchzeri* Hoffman.

in all, there seems to be no good reason to doubt that the floras in both localities are essentially the same.

In Illinois the concretions occur in shales associated with "Coal No. 2," and they are therefore in the Carbondale formation. In Iowa they come from a shale which Professor L. M. Cline has written us is most probably located stratigraphically near the horizon of the Munterville cyclothem. According to the 1942 report of Weller, Wanless, Cline, and Stookey and the 1944 correlation chart of the Pennsylvanian formations of North America by Moore *et al.*, there is, as might be expected, no great difference in the age of the containing strata in these two localities, some 260 miles apart. This fact is, of course, not surprising, for it must have required a peculiar set of conditions for such concretions to form; and in two parts of one general sedimentary basin it is more likely that these conditions were fulfilled more or less contemporaneously than at different times.

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AN OUTCROP OF EOCENE SEDIMENT ON THE CONTINENTAL SLOPE[1]

JOHN NORTHROP AND BRUCE C. HEEZEN

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During August, 1947, the authors carried out submarine geological work on the continental shelf and slope south of Cape Cod, Massachusetts. Exploration was done[2] with the Ewing underwater camera (Ewing, Vine, and Worzel, 1946*a*) modified to obtain a short core simultaneously with the bottom photograph. Five separate profiles crossing the continental shelf and part of the continental slope were completed. The first of these profiles (fig. 1) extended 100 miles offshore due south of Vineyard Lightship and consisted of fifty-three bottom photograph stations spaced about 2 miles apart. At thirty-six of these stations, cores averaging 23 cm. long were obtained from the seaward portion of the profile, where bottom deposits were "soft" enough for the coring tube to penetrate. The cores, except the one obtained at the farthest seaward station, D-20 of the first profile, contained reworked Pleistocene and Recent material.

This paper concerns station D-20 located on the continental slope at 39°50′ N. lat., 70°50′ W. long., at which place the coring tube brought up Eocene Globigerina ooze from a depth of 500 fathoms (1,000 meters wire sounding). The ocean-bottom photograph taken simultaneously shows the Eocene outcrop to be a part of a submarine slope. Similar Eocene material has been reported by Stetson (1949) on the continental slope beneath 10 cm. of Recent "slumped" material in one place and 122 cm. in another. Another occurrence, under 8 cm. of Recent

[1] Lamont Geological Observatory Contr. 20; Woods Hole Oceanographic Institution Contr. 523. Manuscript received June 10, 1950.

[2] Work done on the research vessel "Balanus" of the Woods Hole Oceanographic Institution and supported by the U.S. Navy Bureau of Ships under contract No. NObsr-39092.

material, is also reported from the margin of the Blake Plateau.[3]

The Ewing underwater camera was attached to a 12-foot wooden pole weighted at the base with 10 pounds of lead and fitted with a 1½-inch I.D. steel coring tube, 23 cm. long. This assembly was lowered to the sea floor on $\frac{3}{16}$-inch wire rope paid out from a winch. The apparatus was designed so that, as the coring tube penetrated bottom deposits, a magnet was moved past a magnetic switch, which closed the circuit from the battery to the flash bulb and camera shutter; the core and photograph were thus taken simultaneously. The coring tube usually came up filled with sediment, but at station D-20 it was only three-quarters full, showing that the material was more consolidated than usual and that the Eocene sediment was exposed on the sea floor, as no covering material could have been pushed out of the top of the tube.

The sediment was cream-yellow when wet, but hardened and became chalk-white after drying. Samples from the top, middle, and bottom of the core, respectively, were washed through a 0.061-mm. sieve to catch the microfossils. The coarse fraction contained abundant Foraminifera, the majority of which were *Globigerina*. Samples were submitted to Dr. S. K. Fox, of Princeton University, who states in a personal communication that

> the faunal assemblage is distinctly Tertiary in aspect, probably upper Eocene, closest to the Jackson fauna and maybe dips down a little bit to the Lower Eocene. Some Wilcox and Claiborne elements are present and the material is slightly older than that described by Cushman from Stetson's cores. Some of the forms appear to be more damaged than others and it may represent a slightly reworked deposit. Glauconite is present as casts of some Eocene species.

Dr. Fox has identified the following forms:

A. Those similar to the Upper Eocene fauna of the Jackson formation:

Dentalina jacksonensis (Cushman and Applin)

Robulus alato-limbatus (Gumbel)

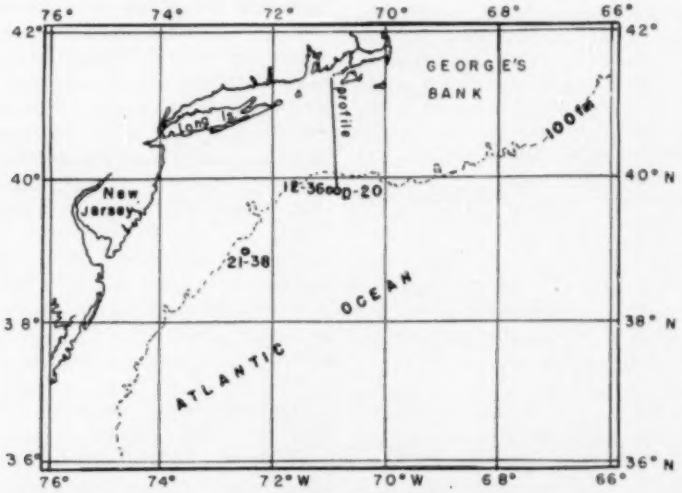

FIG. 1.—Location of profile across the continental shelf and out to the camera station D-20. Stations 12–36 and 21–38 are where Stetson has cored similar material.

[3] Unpublished data of D. B. Ericson describing material in a piston core obtained by the research vessel "Atlantis" in May, 1949.

Eponides cocoaensis Cushman
Anomalina bilateralis Cushman
Anomalina jacksonensis (Cushman and Applin)
Vulvulina advena Cushman
Bulimina elongata d'Orbigny
Nonion micrum Cole

B. Those similar to the Upper Wilcox and Claiborne groups of the Middle Eocene:
Globigerina triloculinoides Plummer
Bulimina ovata d'Orbigny
Parrella expansa Toulmin
Globerotalia membranacea (Ehrenberg)
Globorotalia wilcoxensis Cushman and Ponton, Var. *acuta* Toulmin, 1941
Gumbelina sp., very close to *venezuelana* Nuttal *Cibicides* sp., similar to *Cibicides praecursorius* (Schwager) from the Middle and Upper Eocene of Alabama, but without the lip and slightly larger

C. Those similar to those described by Cushman (1939) from Stetson's Eocene cores:
Karrariella arenasensis Cushman and Bermudez
Pleurostomella cubensis Cushman and Bermudez

D. B. Ericson, of the Lamont Geological Observatory (Columbia University), has also looked at the fauna and states that "the assemblage is not as rich as that in Stetson's cores and is slightly older."

Foraminifera tests in the ooze give it a high carbonate content; random samples of the material analyzed before sieving contain 74 per cent carbonate by weight. Thermal analysis was made of the fraction finer than 0.061 mm. in two samples. Dr. J. L. Kulp (of Lamont Geological Observatory [Columbia University]) says, "comparing the thermal curve with those for carbonates of known chemical composition, the sample contains 40% ± 10% of very finely divided calcium carbonate and 10% ± 5% of dolomite. There is no quartz reaction, but silica may be present in other forms along with some clay minerals." The sediment was therefore deposited in a marine environment that was receiving little terrigenous detritus and yet shallow enough so that the carbonate shells were not dissolved.

The photograph (pl. 1) shows the Eocene sediment as it cropped out on the ocean floor. The core was obtained from a point directly below the black rectangular shadow (top-center of the print) at the same moment that the photograph was made. The camera, axis nearly vertical, was 7 feet above the shadow. About 35 square feet of surface would have shown in the photograph if the sea floor had been horizontal. However, the distinct shadows, which run diagonally across the photograph, show the rock farthest from the camera pole to be lower than that cored directly below the camera. The outcrop is thus in the form of an escarpment in which there are benches of more resistant strata that make the shadows appearing in the photograph. The most resistant layer appears as a series of angular protrusions that crop out in a line from the central left-hand margin of the photograph to the upper right corner.

The authors were unable to map the topography in detail at station D-20 because the "Balanus' " fathometer was inoperative at that depth. However, a chart published by Veatch and Smith (1939, pl. 1) shows a small submarine canyon to be present at camera station D-20. Stetson's cores 12–36 and 21–38 were also taken from canyons. The few centimeters of Recent green silt covering the Eocene in Stetson's cores show that the later continental-shelf sediment has been eroded locally. The outcrop photographed shows that this erosion is occurring today.

ACKNOWLEDGMENT.—Acknowledgment is made to Maurice Ewing, who supervised the research; to S. K. Fox and D. B. Ericson, who identified the Foraminifera; to J. I. Ewing and J. L. Worzel, who aided and technically advised the authors; and to H. C. Stetson and G. P. Woollard and the Woods Hole Oceanographic Institution.

PLATE 1

Ocean-bottom photograph of Eocene sediment outcropping on the continental slope. Station 20-D: depth, 1,000 meters; latitude, 39°50′ N.; longitude, 70°50′ W.; flash bulb No. 5; distance from camera to light, 43 inches; camera to bottom, 77 inches; focus, 4.5 feet; stop, *f*/9; shutter speed, $\frac{1}{25}$ second.

Eocene sediment outcropping on the continental slope

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AN OBSERVATION ON WIND-BLOWN SILT[1]

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A paper by Russell (1944) has again stimulated discussions on the origin of loess. After thoroughly reviewing the numerous and varied hypotheses proposed, he suggested a fluvial-colluvial origin for loess along the lower Mississippi Valley and rejected the more commonly accepted hypothesis of eolian origin. Holmes (1944), Leighton and Willman (1949), and Thwaites (1944) have supported the hypothesis of eolian origin for the lower Mississippi Valley loess, and Swineford and Frye (1945) showed that, contrary to Russell's beliefs (p. 24), wind could sort material to the degree represented by some loess deposits.

The hypothesis of eolian origin, which has been generally accepted for the last thirty to forty years, is that fine glacial debris, spread over the relatively broad vegetation-free glacial outwash plains and floodplains, is transported by winds and deposited on adjacent uplands. This phenomenon can be observed today on some floodplains of glacial streams (Tarr and Martin, 1913, pp. 299–300; Tuck, 1938, pp. 651–653; Rockie, 1946, pp. 5, 10) but is restricted largely to the summer months. In winter, ice and snow cover much of the silt, decreasing the possibility of wind transportation.

[1] Published by permission of the director, U.S. Geological Survey. Manuscript received July 20, 1950.

In central Alaska, bluffs along part of the Tanana and Yukon rivers are blanketed with a well-sorted tan silt, whose origin has been discussed for many years. In connection with studies by the U.S. Geological Survey dealing with the character, distribution, and origin of this silt and other unconsolidated sediments in the Tanana Valley, the writer was able to obtain, during the summer of 1949, some quantitative observations of silt transported by wind from the floodplain of the Delta River.

The Delta River, approximately 85 miles long, is one of many glacial streams that flow northward from the Alaska Range to the Tanana River. The upper part of the river flows in a canyon across the Alaska Range; the lower part flows across the broad Tanana lowland. During the Pleistocene the valley served as one of the main channels for ice moving northward from the Alaska Range across the lowland almost to the Tanana River. At present the river still receives meltwater from many glaciers, which are about 30–50 miles upstream from its mouth, and consequently is heavily laden with silt. Where the river flows across the Tanana lowland, the gradient is relatively low; the floodplain is 1–2 miles wide, with the intricately braided channel and numerous silt-covered bars that are characteristic of glacial-drainage streams.

Moderate to high-velocity northward-

blowing winds from the Delta River canyon in the Alaska Range commonly sweep across the Tanana lowland in the vicinity of Big Delta. Winds up to 25 and 35 miles per hour are common, and velocities up to 97 miles per hour have been recorded (oral communication, Mr. K. Kulm, August 16, 1949) by the Civil Aeronautics Administration at Big Delta airfield, which is adjacent to the river near the junction of the Alaska and Richardson highways.

At 1:00 P.M. on August 20, 1949, when a 14-mile-per-hour wind from the southeast was recorded at the airfield, a cloud of floodplain silt was raised about 50–100 feet above the ground. One hour later the wind had increased to 30 miles per hour and was transporting great clouds of grayish dust from the

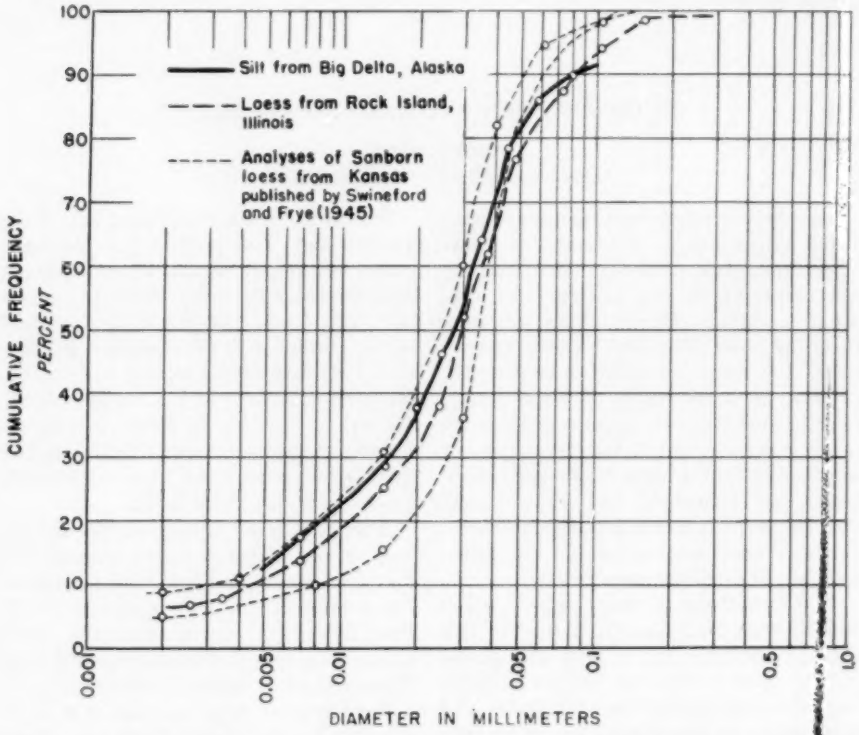

FIG. 1.—Cumulative curves of mechanical analyses of silt adjacent to Delta River floodplain near Big Delta, Alaska; loess from Rock Island, Illinois (17th Street and 27th Avenue); and Sanborn loess from Kansas. Alaskan and Illinoian samples collected by T. L. Péwé and analyzed by the Corps of Engineers, U.S. Army, Rock Island, Illinois. Sanborn loess analyses of samples *1* and *10* from Swineford and Frye (1945, p. 252).

PLATE 1

A, Oblique aerial photograph of clouds of silt transported by wind from the Delta River floodplain, central Alaska. View looking north. Photograph by U.S. Navy, 1948.

B, Same as *A* except that photograph was taken from a lower altitude and view is to the west. Photograph by U.S. Navy, 1948.

A

B

Wind-blown silt, central Alaska

floodplain. The writer noted from an airplane that the silt was carried to a relatively even height of 5,000 feet above sea level, or 4,000 feet above the ground, and was suspended over an area of about 300 square miles. The dust was so thick over this area that, from an elevation of 6,000 feet, it almost canceled the ground below. For the most part, the clouds of dust covered the area west of the Delta River to Delta Creek and south of the Tanana River; however, a few square miles north of the Tanana were blanketed, and some of the silt was deposited on the upland, where silt is at least 10 feet thick.

Unfortunately, none of the dust could be collected in the air; however, silt adjacent to the floodplain on the east was analyzed for size-grade distribution. Comparison of the cumulative-frequency curve of this silt with curves from the analyses of loess from Illinois and Kansas (fig. 1) reveal close similarity in shape. The cumulative-frequency curve of the Alaska silt is also similar to cumulative-frequency curves of wind-blown dust published by Swineford and Frye (1945, p. 252) and Zeuner (1949, p. 27).

Excellent aerial photographs of a similar phenomenon were taken by the U.S. Navy in the summer of 1948 (see pl. 1, *A* and *B*). These photographs illustrate the type of dust clouds observed in 1949; however, the photographed clouds are less extensive than those reported by the writer.

This observation supports the assumption that large quantities of silt are transported from floodplains of glacial-drainage streams and deposited on adjacent flats and uplands. The velocity of 30 miles per hour compares favorably with the velocities of dust-bearing winds cited by Bryan (1927, pp. 39–40). Mechanical analysis of deposited silt compares well with analyses of wind-blown dust and loess from Illinois and Kansas. The size-grade distribution fulfils Russell's requirement for loess, that at least 50 per cent of the distribution must fall between 0.01 and 0.05 mm. (Russell, 1944, pp. 4, 5).

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REVIEWS

Elements of Optical Mineralogy, Pt. II: *Descriptions of Minerals (with Special Reference to Their Optical and Microscopical Characters)*. By ALEXANDER N. WINCHELL, with collaboration of HORACE WINCHELL. 4th ed. New York: John Wiley & Sons, Inc., 1951. Pp. xvi + 551; figs. 427. $12.50.

This edition, the first since 1933 (Fisher, 1934) represents an increase of 90 pages, 65 figures, and 6 dollars. There are ten chapters, as in the earlier edition; the one on organic minerals (vi) replaces that on columbates and titanates (ix). The details of classification show many changes and much improvement; one is reminded of Strunz (1949) and Palache, Berman, and Frondel (1944). There are some hundred and twenty variation diagrams (properties versus composition)—double the number of the previous edition. Data regarding the unit cell and space group are given; spacings from X-ray "powder" films are not. Most changes are hidden in the mass of details that characterize a descriptive mineralogy. The book is a "must" for the serious student and professional worker in optical mineralogy; and all of us should thank the authors and publishers for the work, even though we may regret what inflation has done to the price.

We shall always (no doubt) be plagued by the question: What is a mineral? This is discussed on page vi. Orthoclase and sanidine are treated as separate minerals, but plagioclase is a single one, although it certainly should be two or more. Because Winchell believes that in the spinellids Fe''' can proxy for Cr-Al only to a limited extent, he has but two minerals: spinel (which includes chromite) and magnetite. Quite a different treatment appears in Palache, Berman, and Frondel (1944). "Brownspar" means all the rhombohedral carbonates except calcite and dolomite. "Hornblende" stands for any monoclinic amphibole (p. 425), including "common augite," there is a "polyaugite" (p. 404). Attempts to simplify are in some cases not justified but in most are at least commendable. "Pollucite" (fig. 140) is used for the name of a "molecule" rather than for the mineral. "Melanterite" means an iron salt to most of us, but it is a monoclinic $A''SO_4 \cdot 7H_2O$ in this book.

When the 1933 edition of this work appeared, there was no revised Dana's System, as is presently in process of publication. Thus the citations to the literature will be a less valuable feature of the present edition. But otherwise Dana's System will not outmode Winchell.

The publishers have done their usual fine job; their work is to be highly commended except for the fact that the paper does not take ink well; thus, if one wishes to post corrections, ordinary writing fluid should not be used for this purpose.

Because this book has been so valuable in its field wherever the English language is used, I have spent some time in checking it. The remainder of this review is given over to a listing of the errors or controversial matters noted. This is done solely with the idea that the book has been so important in mineralogical education that the user will want to post some of these corrections to make it more reliable and useful. I have not by any means read all of it, and, in general, I list only those things with which I definitely disagree. The numbers refer to the pages. Where the line number is marked with an asterisk (l*), count up from the bottom of the page, but do not count footnote lines.

20.—Palache, Berman, and Frondel (1944) discuss meteorites under the iron-group metallic elements. Moissanite and osbornite, which are nonmetallic, are covered here, as is siderazot, which has not been found in meteorites. The reason for these anomalies seems to be to avoid setting up a new class of carbides, nitrides, and phosphides for these few representatives. Winchell puts moissanite under metallic elements "because it has metallic bonds between its atoms." This is a remarkable error; it would be far preferable to place moissanite with wurtzite (p. 43) under sulfides.

24, l. 5.—Structure, not space lattice.

30.—Atacamite is dipyramidal.

34.—Reference 42 is to Bull. 833 (not 883).

39, l. 9.—The statement "most sulfides have ionic binding" is an unjustified oversimplification.

46, fig. 28.—This figure is wrong, since it is taken over from the last edition (p. 22) without

change, and yet new elements of crystallization have been employed. It should be redrawn with the proper β-angle and with $X \wedge c = +11°$ (because in reorientation the direction of the $+b$-axis is reversed). It is handled correctly in Palache, Berman, and Frondel (1944, p. 257), where $X \wedge c$ is given as $-11°$, but unfortunately a $(-)$ angle is (apparently) used if the given indicatrix axis lies in acute β, whereas Winchell (p. xv) has most commendably followed Schuster's rules (which have been generally accepted for the feldspars) for all nonorthogonal minerals.

46, fig. 29.—The symbol (113) *should read* (111).

51, fig. 32.—This figure fits the old elements; with the new elements, the face labeled (110) *should be* (101) and (011) *should be* (210).

53.—Enargite should be listed under Division 2 (p. 54). Figure 35 has incorrect curves for N_o and B (see description).

54.—Hutchinsonite belongs under Division 3 (p. 55).

56, l. 7.—Change AX_2 to A_2X.

60.—Winchell has followed Palache, Berman, and Frondel (1944, p. 552) as regards braunite. It seems more probable that it is dimorphous with långbanite (see formula on p. 529) and belongs with the nesosilicates. It is probably $I4c2$ with $a = 9.40$, $c = 18.70$ and U.C. = 8 (Byström and Mason, 1943).

70, figs. 42–43.—These are taken without change from the last edition (p. 59), and yet new elements are adopted which interchange the directions of [*a*] and [*b*]. The description correctly states that $Z = b$.

77, fig. 45.—The 2V angle *should read* 160°+, not 60° ±; the axes (or faces) also need labeling.

89, fig. 56.—This old figure fails to fit the new elements.

93.—Sapphirine may be a nesosilicate (p. 525 or 529).

95.—The element columbium should be called "niobium," according to current agreements. If the symbol for yttrium (e.g., see fergusonite, p. 94) could be abbreviated Yt (instead of Y), then confusion would perhaps be avoided in referring to a silicate as a W X Y Z O structure (Berman, 1937, p. 354).

97.—Blomstrandinite is called "priorite" in Palache, Berman, and Frondel (1944), whereas blomstrandite is "betafite" (p. 94).

116, fig. 75.—This figure must have been inspired by Loupekine's work (see nn. 8*a* and 18). The ordinate should be labeled N_B or $N_{B'}$; the abscissa N_O. There are many other places where additions made apparently in a late manuscript stage are not well integrated.

117, l. 4.—The (010) cleavage listed does not fit the descriptions for strontianite and cerussite.

124, l. 17.—Replace* (101) *by* (102).

127, l. 7.—The sign of extinction is (+) according to figure 85. I would prefer to state $Z \wedge c = -66.5°$—i.e., always to measure an extinction angle against the larger optic symmetry (indicatrix) axis in the plane in question (cf. p. xv).

131, l. 5.—Coal is no more a mineral than is granite. Ordinary banded coal is a residual sedimentary rock that may contain several *rock* units, such as vitrain, clarain, durain, and fusain. Each of these is made up of more or less homogeneous units (e.g., under the optical microscope) which should be regarded as minerals. It is strange that petroleums are not listed in this chapter; elsewhere natural water is properly regarded as a mineral.

The introductory note to chapter vi might lead inexperienced students to think that all carbon compounds are listed in chapters v or vi; but, of course, there are carbono-sulfates, -hydroxides, -phosphates, and -silicates, as well as moissanite and wiserite (p. 90).

135.—It seems illogical to have a separate chapter for borates; why not put them under chapter v, "Carbonates, etc.," along with nitrates and iodates? Also, why not place selenites and tellurites (p. 177) as well as arsenites and vanadites (pp. 236–237) here?

136.—Eremeyevite (preferred name according to Hey, 1950, pp. 66, 379) is misspelled.

139, l. 7.*—U.C. is 4, not 2.

143, l. 11.—The (101) *should read* (102).

143, l. 14.—The extinction angle should be $(-)$; it is better stated as $Y \wedge c = +19.5°$.

150, l. 6.*—The (210) heading *should be* (110).

151–152, figs. 94, 95.—The (010) *should be* (100).

160.—The γ-value given for chalcanthite proper is just 2° too small. The **a**-values listed for all the chalcanthites are the supplements of the correct figures for the Barker orientation. This error seems to have been introduced in Winchell (1931, n. 35, p. 233) under the mistaken assumption that all the interaxial angles could be obtuse.

162.—The old orientation given for polyhalite should be supplanted by the one that

brings out its isomorphous relationship with leightonite (the reference appears in n. 66).

189.—Under fillowite the "albitic granite" of Branchville should be called "granitic pegmatite"; cf. also reddingite (p. 190), fairfieldite (p. 191), and dickinsonite (p. 214).

190.—The "ore deposit" occurrence of salmonsite is a pegmatite.

209, fig. 118.—*Replace* (100) *by* (010).

212, fig. 119.—The slope of the right portion of the 2V curve is not sufficiently steep.

212.—The 2V value of the second line of the table should be 100°, not 80° as given.

236.—Dixenite may be a nesosilicate (p. 529).

241.—The use of atoms (rather than ions) is bad, since $Si_9Al_2O_{24}$ is neutral if it is composed of atoms.

241, l. 8.*—*Read* silica *in place of* silicon.

243–244.—A few diagrams are needed to elucidate these four tables.

245.—The splitting of E 1 (*a*) and (*b*) into anhydrous and hydrous units is shown on page xii.

246.—There should be a *P-T* diagram to illustrate the description given on the upper half of this page.

251.—Footnote 23 belongs on page 242 (with a discussion on B like that given for Be).

253.—Pollucite should be treated as isometric (like analcite), especially since it is isotropic and only the *a*-dimension of the unit cell is given. Line 3 under pollucite, figure 138 *should read* figure 140.

258, fig. 145.—This figure is taken unaltered from the third edition, and yet the *a*- and *b*-axes of danburite have been interchanged.

260.—Banalsite is said to be a feldspathoid (this last term is not in the Index); but see the footnote on page 261. Since banalsite corresponds to a 2:1 carnegieite-celsian compound, chemically it is a feldspathoid and should be on page 257, following nepheline. In fact if the orthorhombic banalsite unit cell be regarded as pseudo-hexagonal, it is nearly identical with that of nepheline except for the doubled *c*-unit; compare low and high tridymite. It should be noted that pseudo-hexagonal symmetry is not apparent in the Laue photograph along [*c*] of banalsite published by Smith *et al.* (1944). The other feldspathoids (besides leucite-nepheline) are on pages 348–354 (separated by scapolite), after the feldspars and zeolites, and on pages 473–475. In short, the feldspathoids are a "genetic group," as are the pseudo-zeolites (datolite, p. 355; prehnite, p. 359; apophyllite, p. 394; and pectolite, p. 460); thus the members of neither of these can be closely associated when a strictly crystallochemical classification is used.

260.—The perfect basal cleavage of petalite would make one suspect it of being a phyllosilicate; even its name indicates this. The good cleavage, called (201), *should be* (101). This error, carried over from the previous edition, has apparently crept into the technical literature (Nel, 1946, p. 52).

The discussion of the feldspars is relatively disappointing, since it shows little revision except for the addition of a section (pp. 317–329) on the Nieuwenkamp method of plagioclase determination without the U-stage. Since the charts required in this technique are reproduced on a small scale (some are omitted), it will be necessary to consult the readily available original publication for accurate determinations; a Berek compensator or similar device is also needed. Those who like this method will be interested in the charts by Laffitte (1950).

The section on the feldspars fails to cover some items of importance and contains many errors. There is no satisfactory discussion of the optics of high-temperature plagioclase; references appear in Köhler (1949), and migration curves are given by Oftedahl (1948, p. 13, data from Tertsch, 1942). The question of order-disorder phenomena in the feldspars, conveniently summarized by Buerger (1948, p. 118), with references, should appear in the section of pages 297–302 to help explain some of the relationships there discussed.

In figure 149, why was the stability line changed near the lower left corner to the position shown? How reconcile this with figure 189? This latter is entirely unsatisfactory (cf. Oftedahl, 1948, p. 68). (A diagram published too late for inclusion is given by Bowen and Tuttle, 1950, p. 497). In the table on page 265, the 001 ∧ 010 values for the plagioclases are the supplements of the interfacial angles. Also, many of the values given in this table conflict with those which appear on pages 308–312. In the discussion of twinning near the base of page 269, why omit references to the work of Donnay and of Emmons and Gates?

The data of figure 166 are seventy-odd years old and are too simplified; modern values could be incorporated as a migration curve in figure 171. This latter figure is not on (010), but on a plane normal to [*b*]; on it the signs of the pina-

coid symbols are reversed; thus (100) *should be* (100), and (001) *should be* (001). In short, the trace of +[*c*] extends vertically *downward*. Moreover, the "Trace of Pericline Ab" is really the trace of pericline An_{13}; the former should be about 24° clockwise (to fit the source). Similarly, in figure 172 the signs of the indices of the face-poles are reversed. The plane (241) listed in the first line of page 275 is obviously wrong (from inspection of fig. 170); it turns out to be (481).

Adularia is said to invert to sanidine at 1,100° on page 274, at 900° on page 276, at 900°–1,100° on page 297, and at 900°–1,150° on page 305. It is shown at 900° in figure 189. In short, this change is a sluggish one; other factors besides temperature are important.

281.—Tsuboi's dispersion curves for $N_{X'}$ of the cleavage flakes of the plagioclases drawn on Hartmann dispersion paper appear as lines sloping gently down to the left in figure 174 (note that the "N_F–N_C" along the left edge of the chart should be raised about 1½ inches). Where these cut the 589.3 line establishes points representing $N_{X'}$ values that duplicate the useful data of figure 173; the latter is therefore unnecessary. The reviewer's students who have used figure 174 find that it does not give satisfactory results on feldspar determination (checking against U-stage methods). Whether this is caused by failure of the grains to lie so that the cleavage is really parallel to the microscope stage, by lack of temperature control, or by other factors is unknown. What experiences have others had?

298, l. 4.—Here it is stated that "less than 1% CaO has ever been found in celsian," but on page 307 a variety with 4 per cent is described. An anorthite with 5.7 per cent BaO is described by Nockolds and Zies (1933, p. 452).

299, ll. 8 and 19.—The albite laminae of perthite are mistakenly stated to lack albite twinning. Both figures 190 and 192 probably refer to cryptoperthites (see p. 311, below).

303.—In the last edition adularia and sanidine were treated as varieties of orthoclase; in this edition orthoclase (including the variety adularia) and sanidine are counted as different species. This is a backward step which is out of line with much of the text (taken unchanged from the last edition). I suggest a return to the usage of the last edition, in spite of the fact that this is somewhat confusing in that the term "adularia" originally meant a certain crystal habit occurring in hydrothermal (not pegmatite; see n. 84) deposits.

311.—The "anorthoclase" of the rhomb porphyries is cryptoperthite (Oftedahl, 1948, pp. 9–10, 56). This and antiperthite are not mentioned; neither is cleavelandite.

312, l. 25.—U.C. is 8 (not 4) for the plagioclases with $c = 14.16$.

330.—There are 29 minerals listed under zeolites (only 12 are shown in fig. 223), of which 3 (arduinite, laubanite, pseudomesolite) are doubtful and 1 (armenite) is not a zeolite. There are no groupings indicated among these (although the first 6 except ashcroftine are non-fibrous), which makes them rather indigestible; the arrangement is in order of decreasing symmetry. Nothing is said about ion-exchanging properties (except a sentence under laumontite). Numerous data on page 332 (especially fig. 224 and the elongation lists) are inconsistent with what appears in the later descriptions. What does the following sentence (p. 332) mean: "Both [heulandite and brewsterite] have the positive acute bisectrix normal to the cleavage whose direction is therefore negative"? In figure 230 the melatopes are wrong; in figure 234 angle β should be 120°; the unit cell of heulandite does not contain 5 of the formulas shown (p. 347).

The "clay" minerals are unfortunately badly scattered; this confusion is not necessitated by the classification scheme, since Strunz keeps them together quite well. Illite is not even mentioned, and ion exchange or the results of differential thermal analysis are not discussed. The kaolinite group (p. 362) has some of its members (anauxite, halloysite) on page 400 as if they belonged with the montmorillonites, and allophane is way at the end of the book (p. 531). That one should not adhere to any arbitrary scheme of classification too rigidly has been recognized in the case of saponite (p. 399, n. 156). Just as opal and lechatelierite, though amorphous, are grouped under tectosilicates near quartz, so allophane could be put under phyllosilicates with the clay minerals.

361.—Pyrophyllite should not be separated from talc by the kaolin minerals.

362, l. 5.—U.C. is 1, not 2.

363, fig. 249.—$Y \wedge a = 23°$ *ca.* (not 11°) if $X \wedge c = +16°$; also 100 *should be* 100; also 2V data do not fit the text.

366.—Even though the author prefers to cling to his classification of the micas, Stevens' (1946, p. 103) classification should at least be described.

370, l. 10.—Parallel with 010 (not 001).

379, l. 4.—U.C. is 4, not 2 (but note that re-

cent work has indicated a change in the U.C. dimensions and orientation given).

381, l. 2.—c = 3.08 (not 1.31) to fit the U.C. dimensions given (p. 382).

383, fig. 262.—This does not show $\beta = 97°$.

386, fig. 268.—Add subscript 2 to Si in formula of antigorite.

424.—This table is a repetition of that on p. 404.

425, l. 2.*—*Change* 100 *to* 010.

426.—In the table the 2V value for the Falun anthophyllite must be measured over Z (not X), according to the indices.

440, fig. 331.—The melatopes are wrong.

445, l. 3–4.*—This statement is untrue. However, for epidote, $2a$ (sin β) = 16.18, which = ca. a for zoisite.

last edition and has indices which do not fit the new orientation used (cf. n. 40). The same remarks apply to figure 350.

461, fig. 351.—This figure is taken from the last edition and does not fit the X ∧ c = 10° given on page 461. New elements are given for pectolite in this edition.

462.—Schizolite (which may be the same as sérandite, p. 443) is reoriented (n. 54), which is avoided in the case of the very closely related pectolite (n. 46).

463.—Beryl should not be separated (by tourmaline) from cordierite (p. 470).

464, fig. 353.—This is confusing; different sorts of symbols (not just dots) should be used.

465, l. 1.—The formula of tourmaline could perhaps better be written as $3[XAl_6Y_3(OH)_4$-

TABLE 1

RELATION BETWEEN FORSTERITE AND THE HUMITES

MINERAL	UNIT-CELL CONTENTS			a'*		b	c	β
	$Mg(OH, F)_2$		Mg_2SiO_4					
A—forsterite	0	+	4	5.98	5.98 = 4×1.49	4.755	10.26	90
B—chondrodite	2	+	4	7.87	7.44 = 5×1.49	4.733	10.27	109°02′
C—norbergite	4	+	4	8.72	8.72 = 6×1.45	4.70	10.20	90
D—clinohumite	2	+	8	13.68	13.44 = 9×1.49	4.745	10.27	100°50′
E—humite	4	+	12	20.86	20.86 = 14×1.49	4.738	10.23	90

* $a' = a \cdot \sin \beta$.

446, l. 2.—The axial ratio given for zoisite (taken unchanged from the last edition) represents $c:a:b$ of the unit-cell dimensions given.

455.—The a-value in the axial ratio of parawollastonite is only half what it should be to fit the unit-cell dimensions given.

457.—Bustamite axial ratio and unit-cell dimensions do not agree. Computing the former from the latter gives 1.067:1:0.959 (with α and β interchanged), which agrees nicely with the wollastonite values (p. 456). The two are isomorphous, but solid solution is apparently not complete between them. Winchell (base p. 457) seems to think that isomorphism implies complete solid solution. Calcite and rhodochrosite are (completely) isomorphous, but there does not seem to be a continuous solid solution between the two (cf. fig. 67, p. 108). Isomorphism is a goniometric concept, solid solution a physicochemical one; there is no need to confuse the two.

459, fig. 349.—This figure is taken from the

$(BO_3)_3(Si_6O_{18})]$, where X = Na, Ca (and rare K) and Y = the other known cations (including Al but excepting B and Si); cf. what appears in line 12 (which lacks a comma between the Na and Ca). It is regrettable that the formulas appearing elsewhere do not agree with this "master."

473.—Dr. Goldsmith states (personal communication) that figure 367 (in which the data are credited to him) gives a wrong impression, in that there is no such compound as the one listed at the left.

498. l. 2.*—Corrected G. values are tephroite, 4.1; fayalite, 4.14.

505, fig. 400.—This is taken without change from the last edition, and yet new elements are given for axinite.

It has long been known that the humites and olivines are isotypous; in fact, this relationship extends over to triphylite and chrysoberyl. Thus these minerals should be oriented in similar fashion with the twofold axis (of the mono-

clinic members) as *b*. Since Strunz has chosen clinohumite (rather than chondrodite) to follow the $c < a$ rule, this results in the orientation shown in table 1, which is to be contrasted with that in the table on page 512 of Winchell. (In Winchell's table the $c'/2$ for humite and the c' for clinohumite should be interchanged.) While it is true that the $Mg(OH, F)_2 : Mg_2SiO_4$ ratio in the humites is (in the order of B–E of table 1) 1:2, 1:1, 1:4, and 1:3 as given by Winchell (taken from Bragg, 1937, p. 150), if this relationship is shown as is done in table 1, it is then easy to see why a' for D = a' for B plus a' for A; and why a' for E = a' for C plus $2a'$ for A. The explanation given in Bragg based on half the number of Mg ions in the unit cell multiplied by the thickness of a "sheet" of oxygen ions in forsterite (= 1.49° A.) is also indicated in table 1. These "additional" (OH, F) ions lie in (100) planes (orientation of table 1), the cleavage plane of the humites, and cause expansion of the lattice in the direction of a'. Figures 405 and 408 are wrong, since they are taken from the old edition with no change, in spite of the fact that β for these crystals is no longer given as 90°; moreover, the direction taken for the *b*-axis in clinohumite is not even the twofold axis (judging from the axial ratio given). Even figure 407 is wrong, since the *b*- and *a*-axes of the 1933 edition are made the *a*- and *b*-axes of this edition. Similarly, alleghanyite (p. 516) does not have $\beta = 90°$ (Smith *et al.*, 1944).

529.—Långbanite should be above heading (c). Also U.C. = 6 (not 4).

The Index is complete; minor corrections include: glauconite, 377; manganapatite and sodaniter are slightly out of position; montebrasite is misspelled.

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Popular Gemology. By RICHARD M. PEARL. New York: John Wiley & Sons, Inc., 1948. Pp. xii+316; figs. 115. $4.00.

This attractive-appearing, well-illustrated small book is easy to read and should be of interest to amateurs. The first seventy pages are devoted mainly to discussion of the properties of gems. This is followed by descriptions of individual gems arranged according to type of cutting, or whether of silica, of organic origin (amber and jet are called "hydrocarbons"!),

synthetic, or luminiscent. There is a very complete index.

A few quotations follow: "The flaws [in diamond] are usually spots of liquid compounds of carbon which failed to harden as the mineral was formed." "They [ruby and sapphire] are heavy gems, as their high specific gravity indicates." "Among the fairest flowers of the gem universe is tourmaline. . . . Color in single hues and in polychrome, color exquisitely blended and sharply contrasted, color richly streaked and delicately modulated, color usually serene, sometimes glowing, often shy and furtive—evanescent tints which come into view and as quickly disappear or merge into others." "Most of the [rhodolite] crystals are tiny . . . but their lovely color is a delight to any eye fortunate enough to see them." "That amazing optical instrument, the spectroscope (Fig. 47), which reveals the intimate secrets of chemistry. . . ."

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Grundwasserkunde. By W. KOEHNE. 2d ed. Stuttgart: E. Schweizerbart'sche Verlagsbuchhandlung, 1948. Pp. 314; figs. 128.

Virtually every phase of the science of ground-water hydrology is discussed to some extent in this textbook. In addition, part of the book is devoted to ground water and capillary water in relation to the water uses of crops and vegetation and the relation of ground water to soil cultivation, construction, and mining.

The author seems to have accumulated a considerable amount of data on relatively shallow, unconfined ground water, but he does not discuss artesian water in any great detail. He presents some interesting observations on the movement of water vapor into and out of the earth and underground.

In discussing methods of measuring the depth to water in wells, there is a surprising absence of any mention of automatic water-stage recorders. This may be explained, however, by the fact that the author believes there is no need for precision instruments to measure the depth to the water table.

The Thiem formula for determining permeability is discussed, but no mention is made of the Theis nonequilibrium formula or any other discharging-well methods.

In the section on geophysical techniques applied to ground water the author states that the electrical-resistivity method has been used to determine the depth to the table water and the depth to salty water. He also mentions that in some cases the method can be used to determine the rate and direction of flow of ground water.

The Bibliography (29 pages) obviously has not been brought up to date, and, although there are ample illustrations, some of them are crude.

Below is a translation of the Table of Contents together with some notes and comments on several sections of the book.

Foreword.—The first edition (1928) was influenced by O. E. Meinzer, of the United States Geological Survey. In the present edition the author has referred to North American and Dutch literature to a large extent.

Chapter i. Explanation (of terms)

a) Pores in the ground and their relation to ground water.
b) Ground water.
c) Capillary fringe. The author states that the crystallization of ground frost draws water from the zone of capillary water and lowers the water table.
d) Root zone and intermediate zone.
e) Effective porosity and ground-water storage.
f) Gradient, rate of movement, and permeability. Includes a discussion of Darcy's law.
g) Definition of terms. Artesian (flowing) water is distinguished from pseudo-artesian (nonflowing) water. Contains a list of international ground-water terms suggested by Meinzer. Dutch and German ground-water terms are compared.

Chapter ii. Seepage of water into the ground.

a) Rainfall, runoff, evaporation, source of ground water.
b) Shape of the water table in relation to gradient. In cross section, the water table (between two points where it intersects rivers or the sea) is shaped like half an ellipse. A profile of the water table, from a point at some great distance inland, to the seashore, is a parabolic curve.
c) Measurements with the lysimeter. The lysimeter is a container that holds about 1 cubic meter of undisturbed (?) soil and vegetation. It has a drain pipe near the bottom, and it is used to measure runoff, infiltation, transpiration, evaporation, and the formation of dew and frost. Data are given for various crops and soils tested.
d) Rate of seepage into the earth. Discusses seepage behavior in hilly forested areas and meadows and into otherwise impermeable soil through drying cracks. Contains observations on seepage when snow or frost covers the

ground and in areas covered by streets and other man-made developments.
e) Seepage to the water table.
f) Penetration of water vapor into the ground and other subsurface movement of vapor. Discusses earth "breathing," which is said to occur as a result of changes in atmospheric pressure, particularly on hills when stormy winds blow. Also discusses the wandering of water vapor underground from places of high vapor pressure to places of low vapor pressure.

Chapter iii. Ground water as part of the general water supply (of the earth).
a) The balance (equilibrium) in the underground water supply.
b) Procedure of measuring ground-water levels.
c) Annual water-level fluctuations.
d) Long-term water-level fluctuations.
e) The most important relations between ground water and the level of river beds.
f) Direction of flow of ground water.
g) Ground water and the ocean. Includes a discussion of tidal fluctuations in unconfined ground water.

Chapter iv. Significance of geological structures in the process of flow of ground water.
a) Introduction.
b) The formations up to the Tertiary (in Germany). Includes a short discussion of Permian salt deposits.
c) Pleistocene and post-Pleistocene deposits. Includes a considerable amount of information on glacial deposits and postglacial soils of Germany. Discusses residual and transported soils that lie on weak, tilted, and horizontal surfaces in humid regions. Also discusses hardpan and "bleached sand" (in soil), dunes, alluvial soils, swampland (moors), and deposits in dry regions.
d) Tilted strata, faults, joints.
e) Individual aquifers and artesian water, groundwater basins, and ground water that is not in motion.
f) Classification of springs.
g) Ground water and petroleum.

Chapter v. Rate of flow, rate of filtration (permeability), and temperature of ground water.
a) Rate of flow of ground water and its determination.
b) Permeability and its determination. Discusses the Thiem formula.
c) Temperature of ground water. Discusses geothermal gradients and effects of withdrawals in temperature. Following are some of the author's statements: (1) Near younger mountains the gradient is about 1° C. per 34 meters and near older mountains the gradient is about 1° C. per 45 meters. (2) The measured temperature of water discharged from a well varies directly with the rate of discharge. (3) Observation wells ending in shallow aquifers near river banks show a larger temperature fluctuation with increasing river velocity. (4) The temperature variation of shallow ground water is smaller on the northern sides of hills than on the southern sides. (5) Large withdrawals (of shallow ground water) result in a lowering of the temperature of the water. (6) Measurements of temperature in a network of observation wells in a "pseudo-artesian" aquifer revealed that large withdrawals from the aquifer resulted in rising temperatures.

Chapter vi. Quality of ground waters and their protection.
a) General.
b) Material dissolved from overlying formations.
c) Iron and manganese.
d) Nitrate, sulfate, silica, and chloride.
e) Hardness.
f) Natural filtration.
g) Artificial treatment and protection of the quality of ground water.

Chapter vii. Domestic and industrial water supply from ground water.
a) Introduction.
b) Obtaining water from wells.
c) Permissible yield.
d) Examples of large ground-water developments. Gives some pumpage data and descriptions of developments in the following areas: industrial area of western Germany, Berlin, Halle, Württemburg, Munich, Vienna, Prague, Florence, and London.
e) Cone of depression. A distinction is made between the cone in the natural, undisturbed water table and the cone in the water table that exists while pumping is going on.
f) Use of geophysical apparatus in ground-water recovery.

Chapter viii. The water supply of vegetation.
a) Introduction.
b) Growth of plants near unattainably deep parts of the zone of capillary water.
c) Plant growth in dry areas, where the zone of capillary water is deep.
d) Ground water and plant growth in humid areas. Contains a considerable amount of data on various crops, cultivated fields, gardens, marshland, flood plains, and forests.

Chapter ix. Ground water and capillary water in relation to excavations and large-scale construction.

Chapter x. The conflict between mining and ground water.

Chapter xi. Artificial interference with ground water.
a) Introduction.
b) Cultivation of trees, bushes, and soil; wind and frost protection.

c) Influence of hydraulic structures on ground water.
d) Influence of waterworks on ground water.
e) Encroachment of ground water into mines.
f) Encroachment of water into excavations for foundations of large buildings.

Chapter xii. The over-all significance of the science of ground-water hydrology.

It should be borne in mind that most of the ground-water data in this book was collected in Germany, where intensive agricultural, industrial, and municipal developments have been in existence much longer than most similar developments in this country. For this reason, particularly, the book should be of interest as a reference to ground-water geologists and engineers in relatively young countries.

ROBERT SCHNEIDER
U.S. Geological Survey

Handbook of Aerial Mapping and Photogrammetry. By L. G. TROREY. Cambridge: At the University Press, 1950. Pp. xvi+178; figs. 80. $5.00.

This book is primarily a how-to-do-it manual for the professional cartographer. Assuming that the reader is reasonably familiar with map-making procedure and problems, he is led, step by step, through the various special methods involved in the preparation of maps from both vertical and oblique air photos. The emphasis, on the whole, is on those methods requiring a minimum of specialized equipment.

The first chapter is devoted to a very brief review of pertinent aspects of perspective geometry. Chapters ii–iv are concerned with oblique photos. Graphic and other methods of determining height and measuring horizontal and vertical angles are outlined, and information not elsewhere readily available is brought together. Certain of the simpler graphic methods for performing these operations, however, are not included, and some of the explanations seem unduly condensed, leaving the reader a bit perplexed. The construction of perspective grids is included; but, of instrumental methods, only one, the use of the Burns plotter, is discussed. No mention is made of such American instruments as the rectoblique plotter and "universal sketchmaster," which have contributed much to the speedup of mapping from oblique photos.

The remaining chapters of the book are devoted to the uses of vertical photos. Chapters v and vi outline the principles of parallax, stereoscopy, and stereometry, and their application to the measurement of height and the drawing of contours with the parallax bar and comparable instruments. Considerable attention is given to factors affecting the accuracy of this type of work.

Chapter vii outlines procedure for radial-line triangulation. Chapter viii discusses the rectification of photos both by photographic methods and by the use of the camera-lucida type of instrument. Chapter ix gives an over-all discussion of the air-survey operation. Chapter x outlines some of the general principles of the more elaborate stereoscopic plotting machines, and chapter xi is devoted specifically to the multiplex projector. An appendix gives the details of routine mapping procedure by the Royal Canadian Engineers.

A bibliography of nineteen entries is included in the book. It is confined mainly to standard textbooks and manuals, mostly British. With one exception, the equipment illustrated and discussed in the book is British also.

The book will be of interest to the photogrammetrist rather than to the geologist. The former will find it a convenient reference, supplementing standard American treatises and giving a somewhat different approach to certain of the common problems of mapping.

H. T. U. SMITH
University of Kansas

Petrology, Stratigraphy, and Origin of the Triassic Sedimentary Rocks of Connecticut. By PAUL D. KRYNINE. (Connecticut State Geol. and Nat. History Survey Bull. 73.) Hartford, 1950. Pp. 247; pls. 29.

Triassic sedimentary rocks belonging to the Newark series occur in a number of structural basins scattered along the Atlantic coast of North America from Nova Scotia to the Carolinas. Of the several hundred published descriptions dealing with these basins, more than 250 concerns the Connecticut Valley Triassic area of Massachusetts and Connecticut. A long-awaited addition to this formidable list of publications, Krynine's report on the Newark sedimentary rocks in the Connecticut portion of the

Connecticut Valley area, represents the most intensive petrologic examination and the most comprehensive discussion of environmental conditions during the Newark sedimentation that have yet been applied to any of the eastern Triassic basins.

Beginning with a definition of problems, the author lists the broad facts of stratigraphy, structure, and age which have been established by previous investigators in the area. This is followed by a list of remaining problems which are to be attacked: characteristics of Triassic sedimentation, of source area, and of basin of deposition; primary and subsequent structure of the basin; and climate and paleogeography of the Newark epoch. With a consistently inductive presentation, the report uses 96 pages, 21 tables, and 46 illustrations to present the basic data on mineralogy, stratigraphy, and petrology of the sediments; 58 pages and 5 illustrations on the climatic significance of the features in the sediments; and 20 pages and 4 illustrations on the genesis of the basin (structure, sedimentation, and paleogeography).

Krynine estimates the gross lithology of the wedge-shaped block of Connecticut Triassic sediments, which has a maximum thickness of about 14,000 feet, as consisting of 73 per cent coarse clastics (arkose, sandstone, conglomerate, fanglomerate), 13 per cent siltstone, 13 per cent shale, and 1 per cent limestone. The clastic lithologic types are regarded as mixtures of these end-members: coarse arkosic (granitic) debris, 98 per cent composed of quartz, microcline, and sodic plagioclase; fine white and red clay matrix, 98 per cent composed of kaolin, gibbsite, sericite, illite, and hematite; and carbonate cement, which is important only near faults. Twenty-three major mineral species are described, as well as several varieties of zircon, garnet, and tourmaline. The heavy-mineral content averages 1.2 per cent for the entire Newark and comprises as much as 7.7 per cent of certain grade sizes.

Three Triassic formations are defined and named. The oldest formation (New Haven arkose), of fluviatile origin, consists dominantly of feldspathic sandstone, arkose, and conglomerate. The middle formation (Meriden) contains lacustrine and swamp deposits, characterized by thin beds of limestone, dark shales and siltstones, and three lava flows. The upper, or youngest, formation (Portland arkose) is similar to the New Haven arkose. Subdivisions of the formations are made on the basis of heavy-mineral content. Thus the New Haven arkose is subdivided on the relative abundance of garnet and certain metamorphic minerals. Each formation contains finer material in the central and western parts of the basin ("normal" facies) and coarser material along the east side, grading into fanglomerates along the eastern boundary fault ("fault" facies). These facies changes in an east-west direction are not accompanied by any important mineralogic changes, and Krynine neatly uses this fact to demonstrate the identity of arkose and "redstone" facies where the same horizon has been duplicated by faulting (pp. 54–56). Facies changes between southern and central Connecticut, however, are characterized by changes both in coarseness of grain and in the relative abundance of indicolite and epidote and are believed to indicate the existence of two groups of alluvial fans fed by separate streams having separate sources of detrital supply. The constituent grains of all facies in all formations are generally fresh, poorly rounded, little sorted; feldspars are generally more rounded than quartz, and lacustrine deposits more rounded than fluviatile.

The definition of climatic terms, a review of the contrasting theories of Triassic environment, and an evaluation of climatic criteria introduce the chapter on the climate of the Newark epoch. There follows an extended discussion on the genetic significance (1) of arkoses: "The two primary factors in the formation of arkoses are the rate of chemical decay and the time during which this decay is at work"; (2) of red beds: "The color of practically all noncalcareous continental red beds, including those of the Connecticut Triassic, is of primary origin" (from erosion of red soils); and (3) of red soils: "Practically all the red soils are formed at the present time in warm humid climates only." It is concluded that the association of red beds, arkoses, and black shales is most easily understood if it is assumed that the sedimentation occurred on "a humid, tropical savanna in the immediate vicinity of a rugged area." A critical examination of the organic and inorganic features of the Newark sediments finds none that are incompatible with such an environment.

Summarizing the development of the basin, Krynine pictures a trough of deposition, at least 35 miles wide, bounded by a fault scarp on the east and receiving sediments from the streams coming off the scarp. Basin depression accompanied sedimentation, and the two were so re-

lated that the surface of the basin was generally almost flat. Drainage was westward except during the middle Newark, when it was blocked by warping and lava extrusions. Temperatures were uniformly about 80° F., and rainfall was in excess of 50 inches annually, but with a long dry season. Postdepositional tilting and block-faulting have complicated the structure of the basin.

This report is considerably more than just a description of the Connecticut Triassic. Indeed, such description is almost incidental to its other values as (1) an illustration of the effective use of petrologic methods in correlation, in the study of the characteristics of sedimentation and in the determination of depositional environments; (2) a summary of the importance of climate in continental sedimentation, and particularly the interrelations of weathering, time, and rate of erosion in various climates; and (3) a pattern of procedure for the systematic study of continental deposits and sedimentary deposits in general.

Much of the effectiveness of the report arises from its objectivity. It defines the problem under attack and the plan of attack; it deals in specific figures or percentages wherever possible; and it sets up classifications on rock types, grade sizes, angularity, red beds, and arkose so that the reader knows precisely what is meant by a term under discussion even if he does not agree with the classification. The over-all effectiveness of the report is lowered, however, by the exceedingly long, repetitious treatment of climate and genesis and by the glib use of percentages where specificity might well have been tempered with indications of possible variability. Thus some statements ("Red beds form 52 per cent of the section"; "Conglomerates form 9 per cent of the New Haven arkose") suggest that the precision of the percentages as applied to the sediments as a whole is greater than it probably is. With the rapid lateral variations in facies, with the complex faulting, which causes uncertainties as to the true thickness of formations and facies (New Haven arkose near New Haven), and with so much of the sediments, particularly the New Haven arkose, deeply buried, it is difficult to believe that the percentages are as unequivocal as they seem to be. It might be helpful to have more specific information on just how the over-all percentages on lithology and mineralogy were computed. These are but minor details, however, in a report that is certainly one of the important contributions to the study of continental sedimentation and an enjoyable, informative addition to geological literature.

JOHN A. REINEMUND
University of Chicago

A Student's Introduction to Geology (Mainly Physical). By GEORGE MACDONALD DAVIES. London: Thomas Murby & Co., 1949. Pp. v+246; figs. 175. 18s.

According to the Preface, this book is intended for first-year students with some scientific background but no previous knowledge of geology; first-year geography students, who generally take a year of physical geology, were also in mind. Because it is written particularly for students in England, the description are chiefly of successions, outcrops, and scenery in that country, for, as Davies points out, many of the modern introductions to physical geology draw chiefly on American examples.

Although it is probably a factual-type textbook of average modern caliber, several criticisms can be leveled at the volume, none of them, however, as to its accuracy or the range of subject matter presented. It is so condensed and so terse that it reads like lecture notes or a syllabus. However, according to the Preface again, this was the author's intention, for he speaks of one of the needs of students as "textbooks to supplement the spoken word," and apparently considers his book a reference work, supplementary to lectures, laboratory, and field, and goes on to say: "Few students want to wade through turgid or nebulous writing; or to pay a high price for their books. Accordingly, the material has been condensed as much as possible without loss of clarity." Perhaps it has, but the condensation is accompanied by loss of attractiveness in style and by the presupposition of facts and terms on the part of the reader, especially in the early parts of the book. This latter difficulty may be partly due to the order in which the topics are taken up.

After a page and a half on the scope of geology, there is a short, somewhat old-fashioned chapter on cosmical geology; Davies accepts Jeans's tidal theory "or some variant of it" as the only theory of the origin of the earth which appears tenable, failing to mention recent astronomically more acceptable theories (the latest reference given at the end of the chapter is 1929). In the third chapter, "Geologic Time," appears such phrases and sentences as the following: "The sedimentary rocks, even though their total thickness is many miles . . ."; "It is

clear that the deposition of the sedimentary rocks, like the evolution of their faunas, must have taken many millions of years," without explanation of what sedimentary rocks are, and no evidence given of why their deposition must have taken so long, even though in the Preface the author states, " 'Test all things' must be our ideal. . . ." An apt simile in the same paragraph, "The geological record may be not unlike a net—a series of gaps held together by string," must be totally lost on beginning student readers (to whom unconformities are not introduced until 82 pages beyond). There follows a good discussion of the radioactive method of measuring geologic time, but such statements as "pleochroic halos in . . . biotite," "tiny zircon crystals enclosed in biotite," must be meaningless to the student, as must such conclusions as "a basalt of Pliocene age has been found to be 13 million years old."

Some matters are cleared up in the next three satisfactory chapters, "Igneous Rocks," "Sedimentary Rocks," and "Metamorphic Rocks," although it is difficult to see how the student gets any clear idea of basalt transgressing horizons before he knows anything about sedimentary successions. The changing environments to be read from a succession of strata or the evidence for transgressing or regressing seas are not pointed out in the chapter on sedimentary rocks.

The next chapters, "Volcanic Rocks" and "Earthquakes," are satisfactory, but chapter ix, "Dip and Strike," goes into perhaps unwarranted detail for a beginning course intended chiefly for nonprofessional students: the excuse is that many problems and exercises dealing with these matters are included in most practical geology examinations in Great Britain. Nevertheless, this is the one chapter of the book which may make it unique and may be its contribution. The instructor seeking problems in determination of true dip from apparent dip, from three points on a bedding plane, or finding stratum contours and outcrop from true dip, may find this chapter helpful, and parts of it may be useful to assign students who have difficulty in visualizing in three dimensions.

Fourteen chapters, in the following somewhat bewildering order, come next, on folds, faults, "Other Structural Details" (among them unconformities) mountains, section drawing, underground water and springs, rivers, lakes, the sea, ice, wind, heat and frost, and rain. The remaining seven chapters then deal with historical geology: "Introduction to Stratigraphy," "Fossils," "Pre-Cambrian," "Lower Palaeozoic," "Upper Palaeozoic," "Mesozoic," and "Cainozoic." The latter four chapters deal mainly with the places in the British Isles where rocks of a given age outcrop rather than with the story of what went on during the respective eras. Page-long lists of fossils, by genera and species, characteristic of most periods or parts of eras, are given, although it is not clear how the student will identify them. The chapter on fossils contains some excellent drawings and gives a good general summary of the chief types of fossil life.

LOU WILLIAMS PAGE
University of Chicago

A Survey of Weathering Processes and Products. By PARRY REICHE. ("University of New Mexico Publications in Geology," no. 3.) Albuquerque, 1950. Pp. 95; figs. 5. $1.00.

This compilation, originally issued as no. 1 of the "University of New Mexico Publications in Geology," has been revised and reissued as no. 3 in that series. The author has made minor additions, most of which are based on material published in the five years following the appearance of the first edition. The Bibliography has been enlarged from 67 to 88 items.

Students of geology and professional geologists alike should be grateful to Dr. Reiche and the University of New Mexico Press for making this long-out-of-print publication available once more. It remains the best short and reasonably complete survey of weathering processes and products known to the reviewer.

F. J. PETTIJOHN
University of Chicago

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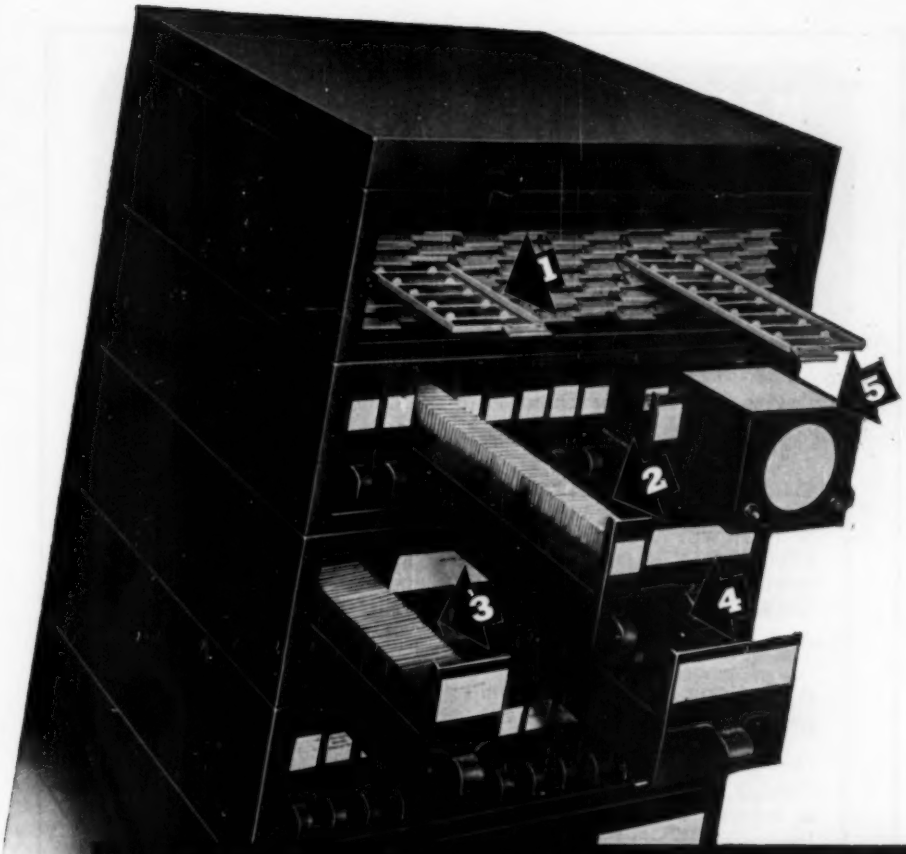

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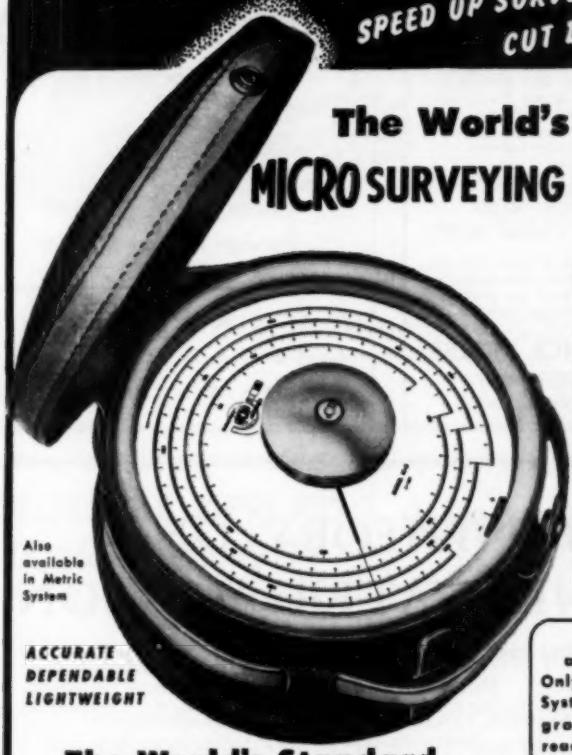

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